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NASA Grant NAG 5-1008: Synthesis of finite displacements and displacements in continental margins

R. C. Speed
M. W. Elison
F. R. Heck
R. M. Russo

Department of Geological Sciences
Northwestern University
Evanston, IL 60208

Scope and Objectives of Project

This is the first semiannual report of this project.

The scope of the project is the analysis of displacement-rate fields in the transitional regions between cratonal and oceanic lithospheres over Phanerozoic time (last 700 ma). Associated goals are an improved understanding of range of widths of major displacement zones; the partition of displacement gradients and rotations with position and depth in such zones; the temporal characteristics of such zones—the steadiness, episodicity, and durations of uniform vs. nonuniform fields; and the mechanisms and controls of the establishment and kinematics of displacement zones.

The objective of such studies is to provide a context of time-averaged kinematic characteristics of the lithosphere for interpretation of NASA's GPS measurements.

Our initial phase of study is divided topically among the methodology of measurement and reduction of displacements in the lithosphere in Appendix 1 and the preliminary analysis from geologic and other data of actual displacement histories from the Cordillera, Appalachians, and southern North America in Appendices 2-4, respectively. Each appended manuscript will eventually be submitted to a journal. The contents of the Appendices are the following:

(NASA-CR-182616) SYNTHESIS OF FINITE DISPLACEMENTS AND DISPLACEMENTS IN CONTINENTAL MARGINS Semiannual Status Report No. 1, 15 Sep. 1987 - 15 Mar. 1988 (Northwestern Univ.) 137 p CSCL 08G G3/46

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Appendix 1: Methodology of measurement and reduction of ancient displacement rate fields in continental margins: M.W. Elison, F. R. Heck, R. M. Russo, and R. C. Speed.

Techniques to determine displacement-rate fields; estimating displacement-rate fields in North American foreland thrust belts; analyses of displacement and strain data; examples of calculated 2-dimensional strain distributions in arbitrary displacement fields.

Appendix 2: Displacement distribution in the cordillera of western North America; M. W. Elison.

Edge of continent; initial structural gradients in the continental margin; structural division is the foreland thrust belt and relation to initial structures; displacement direction, magnitude, and timing in foreland and hinterland; onstrike displacement gradients; displacement transfer zones; foreland width-strain relationships; depth relations and mechanisms.

Appendix 3: Alleghanian displacements in the Appalachian Mountains: F. R. Heck.

Alleghanian tectonics; timing of events; depth-displacement zonations; magnitudes, directions, and rates of displacements; position of craton margin; distribution of Alleghanian strike-slip faults.

Appendix 4: Phanerozoic tectonic evolution of southern North America-Caribbean-northern South America: R. C. Speed.

Event timetable; kinematic regimes; early passive margin kinematics; collision of North America-South America in late Paleozoic time; Mesozoic divergence in Pangea; Gulf rifting; Caribbean-North America; Caribbean-South America; Caribbean oceanic plateau and rifting; oblique collision in South America.

Summary of Findings

1. In the Mesozoic cordilleran foreland thrust belt of the western U.S., the total displacement varies smoothly on strike from 200 km in the north (southern British Columbia) to 100-120 km in the south (Utah). In contrast, the shortening strain varies irregularly on strike, and its magnitude and orientation are related to preexisting structures. Major salients and recesses are not harmonic with lateral variations of total displacement.

2. The duration of motion in the Cordilleran thrust belt varies substantially on strike, as recognized chiefly in age of onset. Whereas the cessation of thrusting is everywhere at 50 to 55 ma, the onset varied by as much as 50 my, between 150 and 100 ma.

3. Mean displacement rates in the Cordilleran thrust belt vary on strike between 1.4 and 2.8 mm/yr. The rates are in fact variable over the duration at each position, implying sporadic thrust movement and probably, intervals of quiescence.

4. A new theory is presented for kinematic compatibility between foreland and hinterland thrusting in the continental margin. The thin-skinned displacements of the foreland are accommodated by ductile horizontal shortening and vertical thickening of crystalline basement in the hinterland. The basement below the foreland (the craton) does not deform and is the buttress against which the suprajacent cover of the thrust belt deforms.

5. In the late Paleozoic Alleghanian foreland thrust belt of the Appalachian Mountains, total displacement diminishes south to north from about 218 km to 138 km. The average displacement direction is N45°W.

6. Strike-slip faults of the Alleghanian orogen occur outboard of the foreland. Displacement magnitudes of such faults range from 12 to 210 km in the northern Appalachians and >17 to 160 km in the southern Appalachians. The average displacement direction is S52°W.

7. Durations of Alleghanian faulting are estimated to be 38 ma for the foreland thrusting and 28 ma for strike slip faulting. The foreland displacements range from 3.6 to 7.1 mm/yr; the principal uncertainty is the application of constant duration on strike through the entire belt. The range of strike slip rates in the northern Appalachians is 0.74 to 5.5 mm/yr and in the southern, >0.63 to 5.0 mm/yr.

8. The apparent restriction of Alleghanian strike-slip faults to the outboard region of the orogen suggests their development was related to deep ductile shearing and the edge of Proterozoic North America.

9. The fundamental shape of southern continental North America was created by intracontinental divergence at the beginning of Paleozoic time. The edge of North America so formed has been the locus of both divergent and convergent kinematic events in subsequent times.

10. The structures created in the multiphase tectonic evolution of southern North America have orientations controlled at least partly by preexisting structures. The effect of inherited discontinuities or anisotropy appears more important in structural orientatons than farfield displacements.

11. Although the Mesozoic divergence between North and South American cratons appears to have included uniform and smoothly varying displacement-rate fields, the intervening region may have undergone complex kinematics, including multiple displacement zones of diverse orientation and rotations.

12. Each of the Phanerozoic major kinematic regimes affecting southern North America appears to have been substantially oblique and included ratios of margin-parallel to margin-normal components ≥ 1.

13. Collisional regimes in southern North America and northern South America have apparently had partitioned displacements between more nearly continuous margin-normal components and more widely discrete margin-parallel or margin-oblique components. Margin-normal displacement gradients may have been largely controlled by ramp effects of the passive continental margins.

14. Each of the major displacement zones of southern North America occupies belts of great width, at least 200 km.

Appendix 1

METHODOLOGY OF MEASUREMENT AND REDUCTION OF ANCIENT DISPLACEMENT RATE FIELDS IN CONTINENTAL MARGINS

M.W.Elison, F.R.Heck, R.M.Russo, and R.C.Speed

Department of Geological Sciences

Northwestern University

Evanston Illinois, 60201

Techniques to determine displacement fields

Fundamental measurements necessary to determine displacement fields are the magnitude and orientation of displacement and the period over which displacement occurs. Given these measurements, time averaged displacement rates and directions can be calculated.

A finite displacement can be estimated if the position of a point prior to displacement and its position after displacement, relative to some reference frame can be determined. Measurements of the initial and final position of a point only give the finite displacement of the point. Because the incremental displacement path is not constrained, the finite displacement will necessarily be a minimum estimate of displacement. The product of a field of general displacements is deformation of which three components are recognized: translation, rotation, and strain. In the case of a single deformation, the current position of a point represents its position after deformation. The necessary information is, thus, the position prior to deformation. Different methods are required to determine the initial position of a point for each deformation component. The methods available are described below. Where more than one deformation component obtains several methods may be required.

Translations are most easily measured when displacement causes offset of initially continuous features such as: isograds; isopachs; sedimentary, igneous, and metamorphic contacts; and geomorphic features. Both horizontal and vertical displacements due to translation can be determined by offset. Holding one portion of the offset feature fixed provides a reference frame for measurement of the relative translation. Offset of initially continuous features allows resolution, depending on the scale of the offset feature, of translations as small as centimeters. Paleomagnetic inclination anomalies may also allow translation estimates. The anomalies only record the latitudinal component of translation, and require knowledge of the rotational and strain history of the sample. The reference frame for paleomagnetic measurement is the Earth's spin axis. Paleomagnetic inclination data may resolve translations of 100's of km but rarely smaller translations.

Rotational displacements can be determined if rotation has produced angular variation in initially planar features. Bedding and foliation can be considered initially planar and, thus, may record rotational displacements. Angular variation of initially linear features may also record rotation. Only some components of the rotation are recorded depending on the orientation of the line. The reference frame for angular variation is relative to some orientation of the feature held fixed. Paleomagnetic declination anomalies provide another measure of rotation, but they only provide the component of rotation about a vertical axis. The Earth's spin axis is the reference frame. Again, the resolution of rotations by angular variation is much greater than the resolution allowed by paleomagnetic data.

Displacements due to strain must be deduced by structural analysis. Strain within rocks results in changes in line length and in the angle between nonparallel lines. Measurement of these changes allows calculation of the finite strain. Derivation of the finite strain provides only the percentage change in line lengths due to deformation. The finite strain must, therefore, be integrated over the region of known strain (either homogeneous strain or a known function of strain with position). Strain measurements provide displacement relative to some point held fixed either within or external to the strained area. The resolution of finite strain is very good, but the determination of areas of constant or continuously varying strain is difficult and decreases the resolution of finite displacement by strain measurement.

Vertical displacements represent a special case in displacement determination because the Earth can be considered to be spherically stratified. Thus, conditions, including sea level and temperature and pressure vary with distance from the center of the Earth. Where sea level or depth are recorded in rocks (i.e., marine terraces or metamorphic grade) the initial position of the rock is known and a reference frame, elevation, is provided. Furthermore, measurements of vertical displacement can be made independent of the displacement mechanism because the initial position is not determined relative to any undeformed correlary. The resolution of vertical displacement depends on the range of depths over which specific recognizable conditions obtain. Thus, marine terraces recording sea level which varies over meters, provide greater resolution than metamorphic grade recording barometric variation over kilometers.

Most deformations in the geological record are combinations of translation, rotation, and strain. Folds and simple shear, for example, are both combinations of rotation and strain. In the common situation of folds developing due to motion on associated faults, rotation, strain, and translation all contribute to the total finite displacement. These situations require combinations of the methods outlined above in determination of displacement magnitude and orientation.

Given that displacement in a region can be estimated, the question is; over what time interval did the displacement occur? Two primary methods of age determination are available: biostratigraphy and radiometric dating. The combination provides the ability to date most rocks. Widely used constraints on the minimum age of a displacement are the ages of igneous bodies that cross-cut and/or stratigraphic intervals that overlap deformation structures. Furthermore, the age of the youngest deformed rocks provides an estimate of the maximum age of displacement. Deposition and igneous activity are rarely continuous, particularly within orogenic belts. Therefore, constraints on displacement period based on igneous and stratigraphic relations commonly give maximum displacement periods. Depositional histories of foreland basins may also provide constraints on the maximum age of displacement. Where sediment within a foreland basin can be directly related to displacing source areas, the age of the sediment records the initiation of source erosion and estimates the initiation of displacement. Finally, dating of widespread periodic features, such as marine terraces and magnetic reversals, in a single locality allows the age of such features from other localities to be dated by correlation. Resolution of ages by biostratigraphy or radiometric dating varies dependent on which fossil or radiogenic element is used. Resolution always decreases with increasing age.

Often displacement orientation and magnitude values are arrived at independent of displacement period values. The time averaged rate of displacement is then calculated from the independent measurements. Because displacement magnitude estimates are generally minima and displacement period estimates are generally maxima, calculated displacement rates are minima. When coupled displacement/age measurements are available the problem of independent measurement can be avoided and displacement rates can be well constrained. A coupled measurement requires the following:

1) the age of a feature must be determined

2) the displacement of that feature, relative to some reference frame, must be determined

3) the age of the feature must lie within the age range over which displacement occurred.

Time versus blocking temperature analysis for metamorphic mineral assemblages represent one example of coupled displacement/age measurement (Dallmeyer and others, 1986; Archibald and others, 1983). Where coupled measurements of two or more features with different ages within the age range of the displacement can be made, the difference in finite displacement over the difference in age gives the displacement rate. If several coupled measurements are available, then changes in displacement rate can be resolved. The measurements that have been made for displacement magnitude, period, and rate and their uncertainties and resolution are discussed in the following pages with specific reference to the western North American Cordillera and to Alleghanian deformation in the eastern United States.

Western North American Cordillera

The method of determining displacement rates that has proven tenable in the fold and thrust belt of the western North American Cordillera involves estimates of horizontal displacement magnitude for individual thrusts and for the thrust belt as a whole coupled with independent estimates of the period over which displacement obtained. In general displacement magnitude estimates are minima, whereas, displacement period estimates are maxima; derived displacement rates are, thus, minima. Various methods are used to determine both displacement magnitude and period and these methods vary in their precision and resolution. Although placing well defined error bars on displacement estimates is difficult, some conclusions regarding the resolution of the methods can be drawn. The following discussion is an analysis of the techniques of determining displacement magnitude and period, problems with those techniques and the effects of problems on the resolution and precision of the derived displacement rates.

The method commonly used to determine displacement magnitude across portions of the fold and thrust belt is palinspastic reconstruction or retrodeformation (Bally, 1984; Price, 1981; Woodward, 1986; Dixon; 1982; Harrison and others, 1980). A detailed cross-section of the fold and thrust belt is constructed and structures are sequentially restored to the pre-deformational configuration. The position of points in the restored section relative to their position in the deformed section provides a measure of the displacement magnitude. This technique has been widely employed in the fold and thrust belt of western North America (Fig. 1; Table 1). Where more than one reconstruction has been attempted, comparison of the displacement magnitudes determined by different studies allows consideration of the precision of the technique. In northern Montana, two reconstructions suggest displacement magnitude across the fold and thrust belt of 170 and 180km respectively (Table 1). In the Wyoming-Idaho-Utah portion of the fold and thrust belt reconstructions vary in displacement estimates from 80km to 140km. Taken at face value, these results indicate that retrodeformation varies in precision from less than 10% to approximately 50%.

Several factors influence displacement magnitude estimates by this method and careful consideration of these factors can improve the resolution of the technique. An obvious consideration is the possibility that some structures are unrecognized and omitted from the deformed cross-section being reconstructed. A correlary problem is the possibility of variable interpretation of structures at depth in the deformed cross-section. These problems point out the importance of well constrained cross-sections in the precision of palinspastic reconstructions.

Unrecognized structures and variable structural interpretation have become less significant problems over time. Because the fold and thrust belt has high potential for petroleum production, closely spaced seismic reflection profiles and wells allow detailed structural geometry to be determined (Dixon, 1982; Woodward, 1986; Perry and others, 1983). Furthermore, geophysical data, including gravity and magnetics can be incorporated to further constrain structural geometries (Harrison and others, 1980; Fountain and McDonough, 1984).

The displacement magnitude of individual thrusts at various positions along strike can be resolved to within a kilometer to a few kilometers (Mudge and Earhart, 1980; Woodward, 1986) suggesting that displacement magnitude for the fold and thrust belt can be resolved to within roughly 10 to 20km, dependent on the number of thrusts within a segment of the belt. Notwithstanding, the ability to constrain structural geometry and produce highly detailed deformed cross-sections, and the good resolution possible for individual thrusts, variations in fold and thrust belt displacement magnitude estimates remain. Bifurcation of thrusts and displacement transfer pose a problem because summation of displacement magnitudes determined on individual thrusts at different positions along the fold and thrust belt may overestimate or underestimate displacement magnitude. Well constrained studies from Wyoming (Woodward, 1986; Dixon, 1982) show that the magnitude of displacement on individual thrusts or thrust systems can vary significantly (Table 1). High displacement magnitudes on one thrust or thrust system often correlate with low displacement magnitudes on others thrusts at the same position (Dixon, 1982; Goldburg, 1984) suggesting displacement transfer between fold and thrust belt structures. The possibility of displacement transfer is illustrated by comparison of variations in individual thrust displacement magnitudes with displacement magnitudes for the fold and thrust belt as a whole. The estimates of displacement magnitude for the entire fold and thrust belt at various positions in western North America show regular north to south variation of approximately 100km or roughly 50%. Individual thrusts can vary in displacement magnitude by a factor of 4 or more (Dixon, 1982; Mudge and Earhart, 1980). It is important, therefore, to determine displacement magnitude at a single position within the fold and thrust belt.

A final consideration in displacement magnitude determination from reconstruction of cross-section is the possibility of large continuum strain in the section. Studies of strain in folds within the fold and thrust belt (Allmendinger, 1982; Gockley, 1985; Johnson, 1985) have shown that locally high strains occur and that the orientation of compression and extension due to strain are not necessarily the same as orientations due to thrusting. However, high strains are localized in the hinge regions of folds (Allmendinger, 1982) and given the high limb to hinge ratio characteristic of fold and thrust belt deformation and the distributed occurrence of folds, the underestimation of fold and thrust belt displacement magnitude due to omission of strain is probably not great.

The regular pattern of displacement magnitude across the fold and thrust belt in western North America and the similarity of estimates produced by different workers at a single position suggest that displacement magnitudes are resolvable to within 10 to 20km or about 10% of the total displacement. The displacement magnitudes are minima due to omission of continuum strain and to the determination of finite displacement as opposed to incremental displacement path. Notwithstanding the problems discussed above, given adequate geological and geophysical data coverage, displacement magnitudes can be better constrained than estimates of displacement period.

Because throughout the western North American Cordillera, thrusts commonly developed from west to east, dating the displacement period of the easternmost and westernmost thrust of the fold and thrust belt provides an estimate of the displacement period for the entire belt. In the eastern part of the fold and thrust belt, foreland basins record the timing of the earliest derivation of material from thrust sheets to the west, the maximum age of thrusts that overrun western portions of the foreland basin, and the minimum age for the cessation of thrusting by strata that overlap the easternmost deformation. In western parts of the fold and thrust belt, the youngest overrun strata are often Paleozoic and synorogenic deposits are uncommon. Based on the sedimentary record, the minimum age of the onset of deformation and the minimum age of the cessation of deformation can be constrained (Table 1). Because displacement period depends on determining the maximum age of displacement onset, as well as the minimum age of cessation, displacement period estimates suffer from poor control in the western parts of the fold and thrust belt. Subsidence of foreland basins due to the load of thrust sheets (Speed and Sleep, 1982; Jordon, 1981), synorogenic conglomerates (Wiltschko and Dorr, 1983; Heller et al, 1986; DeCelles, 1986; DeCelles and others, 1987), and cross-cutting and/or deformed igneous bodies (Mudge and Earhart, 1980; DeCelles, 1986; Harrison et al, 1980) all provide timing constraints for individual thrusts or thrust systems, but generally only constrain the minimum ages for individual thrusts.

The cessation of deformation in the fold and thrust belt is well constrained by overlapping strata, cross-cutting dikes, and radiometric dates on authigenic minerals created by thrust sheet emplacement (Hoffman et al 1976; Aronson and Elliott, 1985). Although the constraints vary from maximum ages of 65my (Thompson, 1979) to minimums of 40my (Bally, 1984; Price, 1981), taken together the data indicate that the fold and thrust belt had ceased activity between 50 and 55my. The onset of deformation is constrained at different positions along the fold and thrust belt to be older than 110 to 150my on the basis of the first orogenic detritus derived from the west and deposited in foreland basins. In Montana and in central Utah, where the oldest orogenic strata are 115my and 110my respectively (DeCelles et al, 1986; Villien and Kligfield, 1986), the orogenic strata overlie an unconformity that locally represents a gap in the stratigraphic record of more than 30my. On the basis of stratigraphy, subsidence, and cross-cutting relations in the eastern part of the fold and thrust belt, the onset of deformation is only constrained to be older than Late Jurassic (165 to 155my) or Late Cretaceous (115 to 100my), depending on location.

In the areas west of the fold and thrust belt, abundant plutons and metamorphic rocks can be dated radiometrically. Radiometric ages are generally precise to within 2 to 5% although interpretations of these ages are dependent on knowledge regarding the thermal history involved. Unfortunately the relationship between the metamorphism, plutonism, and deformation in these western areas and deformation in the fold and thrust well is not well known. Modern models of the relationship (Speed and others, 1988; Brown and others, 1986) consider the shortening of cover strata in the fold and thrust belt to be balanced by shortening of basement rocks to the west. If true, this idea suggests that the timing of deformation to the west should provide a maximum time for the onset of fold and thrust belt deformation. In areas west of the fold and thrust belt and east of the eastern limit of accreted terranes in southern British Columbia, northern Washington and Idaho, and western Utah and Nevada, there is evidence for deformation as old as Mississippian or Devonian and for deformation ranging from Early Triassic to Tertiary. In central Idaho, there is no evidence for deformation, exclusive of accreted terranes, older than Late Cretaceous. Table 2 gives ages of plutons and times of metamorphism for various positions along the Cordillera, west of the fold and thrust belt. Although deformation occurred earlier at some locations, it is interesting that pulses of plutonism and metamorphism associated with deformation are documented in middle to late Jurassic time (175 to 150my) and in Late Cretaceous time (115 to 95my)(Archibald and others, 1983; Fox and others, 1977; Armstrong and others, 1977; Speed and others, 1988; Miller and others, 1988; Snoke and Miller, 1988; Miller and Engels, 1975; Allmendinger and others, 1985; Parrish and Wheeler, 1983; Journeay and Brown, 1986; Armstrong, 1975; Dallmeyer and others, 1986). Metamorphism, plutonism, and deformation occurred in areas to the west at times that correspond to two estimates of displacement onset derived from foreland sedimentation in different locations. Thus, although the metamorphic and plutonic pulses in areas to the west are not directly linked to deformation in the fold and thrust belt, the temporal coincidence of thermal activity and deformation to the west and deformation and foreland basin sedimentation in the fold and thrust belt argues for displacement onset during Late Jurassic time and for a major pulse of deformation during the late Cretaceous.

Table 1 gives the derived displacement rates averaged over the fold and thrust belt and over the displacement period as constrained from foreland basin sedimentation and as estimated from activity farther west. Best fit displacement rates were determined from the average displacement magnitude (where more than one value has been obtained) and the displacement period constrained from foreland basin sedimentation. The uncertainty in the displacement rate was estimated from the minimum displacement magnitude (20% less than the displacement magnitude where only a single estimate is available) and the maximum displacement period allowable given constraints from foreland basins and from within the fold and thrust belt. While the uncertainties in displacement rate derived this way are in some cases large, the displacement rates and their estimated uncertainties demonstrate that 1.0 and 4.0mm/yr are rough limits on the time averaged rate of displacement in the western North America fold and thrust belt. Variations in rates of thrusting over time and as a function of position both along and across strike in the fold and thrust belt and rates of vertical uplift west of the fold and thrust belt based on time versus blocking temperature studies in metamorphic rocks are discussed in the attached manuscript "Displacement Distribution in the North American Cordillera".

Alleghanian displacements in the Appalachian Mountains

In the Appalachian Mountains (Fig. 2), magnitudes of Alleghanian displacement due to northwest-southeast shortening are best known in the Valley and Ridge fold and thrust belt where magnitudes are determined from balanced cross-sections and/or deformed versus undeformed line lengths. Displacement directions are assumed to be parallel to the cross-sections that trend about normal to strike. The magnitudes and directions are given with respect to undeformed North American continent. For example, the Blacksburg, Virginia cross-section (section 3, Fig. 2b) estimates Valley and Ridge shortening between the undeformed continent and the trailing edge of the Pulaski thrust sheet at a point just west of the Blue Ridge-Inner Piedmont thrust sheet. The estimated shortening is the sum of the shortening within the

Pulaski sheet (ranging from 80 to 136 km with an average of 108 km; Bartholomew, 1987) plus the displacement along the Pulaski thrust (100 to 110 km, average 105 km; Bartholomew, 1987) plus the shortening due to folds and thrusts below and west of the Pulaski sheet (33 to 59 km, average 46 km; Kulander and Dean , 1986). The best pick total shortening is therefore 259 km although the amount could vary by as much as 46 km. The displacement direction is taken to be N28W parallel to the cross-section trend which is normal to structural trends at Blacksburg. Although northwest-southeast Alleghanian shortening clearly occurred in the high metamorphic grade eastern Piedmont, no shortening estimates are available because suitable marker horizons are absent and Alleghanian effects are difficult to distinguish from older deformations except where radiometric ages are available, and in those cases shortening estimates have not been done. The same situation applies to the northern Appalachians.

Alleghanian strike-slip displacements occur only in the Piedmont province in the southern and central Appalachians and in the easternmost portions of the northern Appalachians. Magnitudes of strike-slip displacement are determined by the offset of features across the strike-slip zone (eg., plutons, isopach contours, terrane boundaries) or by estimates of simple shear strain versus distance across the zone. Displacement directions are assumed to be parallel to the fault zone or to mineral elongation lineations if they are present. In all cases the magnitudes of strike-slip displacement are more imprecise and poorly constrained than are displacements in the Valley and Ridge because: 1) area balancing is not possible as in Valley and Ridge cross-sections, 2) the types of offset features have boundaries that are generally unsuitable for precise measurements of offset, 3) unrecognized dip-slip motions may be present which could result in strike-slip offsets that are apparent and not real, and 4) displacement estimates by shear strain versus distance across the shear zone invariably assume simple shear alone has occurred with no component of shortening normal to the zone (an invalid assumption in most cases) and the position of shear zone boundaries is often poorly constrained.

Magnitudes and directions of strike-slip displacements are given with respect to a point on the opposite side of the strike-slip zone rather than to undeformed North America. To find the displacement with respect to undeformed North America the displacement vector for the fault in question must be added to all orogen-parallel and orogen-normal displacement vectors between the fault and undeformed North

America (Fig. 2c). However, the uncertainties in magnitude and direction of such a vector are probably quite large because the procedure assumes that all displacement vectors between undeformed North America and the point in question are known. The assumption is acceptable for Valley and Ridge displacements but clearly not for the Piedmont or the northern Appalachians where an unknown amount of northwest shortening occurred in the Alleghanian and where all zones of strike-slip displacement have probably not been identified or if identified their displacements are not always known (eg., the Brevard zone, and others). Therefore, at this time we do not present the strike-slip displacements with respect to a point on undeformed North America although this clearly is an eventual goal of the study.

As an example of strike-slip displacement estimates, the Brookneal shear zone near Brookneal, Virginia (Gates and others, 1986) is about 4 km wide with a shear plane orientation of N40E, 50SE. Shear zone fabrics indicate dextral displacement subparallel to the zone. Displacement magnitude is estimated as $>$ 17km by determining simple shear strain across the zone according to changing cleavage orientations. The displacement magnitude is considered to be a minimum because simple shear is heterogeneous across the zone and all areas of high simple shear have probably not been recognized. Furthermore, the 4 km width of the zone is probably a minimum value. If the zone is transpressive, however, a component of shortening normal to the shear plane will produce the observed cleavage relations at lower simple shear strains which will in turn reduce the estimated strike-slip displacement. Elongate mineral lineations plunge shallowly N50E which when considered with the shear plane orientation and dextral sense of shear indicates a small component of up dip displacement. Furthermore, the Bowen Creek fault zone to the northwest is related to the Brookneal zone and is clearly transpressive (Gates, 1987). Therefore, the Brookneal zone probably experienced at least some component of shortening normal to the zone which has the effect of reducing the minimum displacement estimate (although we have no way of knowing by how much). The displacement direction of rocks outboard of the Brookneal zone is taken as S50W; dextral displacement parallel to the shear zone stretching lineations.

Rates of Alleghanian displacement are more uncertain than magnitudes and directions due to the added uncertainty in the timing of displacement. Displacement timing is reasonably well constrained for individual structures in the Piedmont province and northern Appalachians because cross-cutting intrusives, metamorphic ages, and overlapping strata are common. In the Valley and Ridge however, such features do not occur; timing of deformation cannot be evaluated independently at different positions so the province must be treated as a single package with respect to time interval of deformation. Good estimates of the time of Valley and Ridge deformation are provided by recently published K/AR illitization ages (303 +/-13 Ma to 265 +/-13 Ma) from the Valley and Ridge and Plateau provinces of Alabama, Tennessee, Kentucky, and Virginia (Elliot and Aronson, 1987). These ages are compatible with stratigraphic constraints on timing of deformation. They follow and partly overlap the ages of early Alleghanian clastic wedge strata that were deposited in the Valley and Ridge during uplift to the east and later deformed as Alleghanian displacements migrated westward. Also, the age of the youngest preserved unit affected by Alleghanian deformation, the Dunkard Group (286 +/-12 Ma to 266 +/-17 Ma; Secor and others, 1986a), falls within the time interval of illitization. Deformation of the Dunkard probably represents the waning stages of Alleghanian deformation because it is among the westernmost of deformed strata and is weakly deformed in broad, open, upright folds. Alleghanian deformation therefore probably ended shortly after this interval. A more conservative estimate of the end of Alleghanian deformation in the Valley and Ridge is 235 Ma which is about the youngest age of Alleghanian deformation recognized in the Appalachians (see above). We therefore use 38 Ma (303 - 265 Ma) as a minimum time interval and 81 Ma (316 - 235 Ma) as a maximum time interval of Alleghanian displacement in the Valley and Ridge province.

Table 3 presents displacement rates of Alleghanian strike-slip and fold/thrust structures for which the required displacement magnitude and timing data are available. The displacement rates were determined in the following manner. For each point we choose two values of displacement magnitude and two values of time interval over which displacement occurred. The two displacement magnitude values are the maximum and minimum estimates in the cited reference. If only a single value of displacement was provided in the cited reference we use 25% less and 25% more of the value as minimum and maximum estimates of displacement. The two time interval values are the interval which is the best pick estimate in the cited reference and the interval which equals the best pick plus the uncertainties at each end of the best pick. For example, if the time of displacement is given as between 270 +/-15 Ma and 250 +/-10 Ma the two time interval values we use are 20 Ma (270 - 250 Ma) and 45 Ma (285 - 240 Ma) (see below for how we use them). If the cited reference does not provide uncertainties at each end of the best pick time interval, we add 50% of the best pick as a measure of uncertainty to obtain the second time interval. Where timing of displacement is given by the age of stratal intervals (eg., Pennsylvanian age) we use the DNAG best pick ages (Palmer, 1983) for our best pick interval (320 - 286 = 34 Ma interval for the Pennsylvanian) and add in the uncertainties of the DNAG best picks for a maximum estimate of the displacement time interval (34 + 20 + 12 = 66 Ma). In some cases deformation is estimated to begin between two times and end between two other times. For example, deformation began between 290-280 Ma and ended between 260-250 Ma. In this case a minimum time interval is 20 Ma (280 - 260 Ma) and a maximum interval is 40 Ma (290 - 250). These values are used with the maximum and minimum displacement magnitudes respectively to determine maximum and minimum displacement rates. The best pick time interval for calculating displacement rate is taken as the average time interval or 30 Ma.

The best pick displacement rates in Table 3 are determined by dividing the average displacement by the best pick displacement time interval. In the case where displacement occurred within the best pick interval the uncertainties are given as the difference between the best pick rate and minimum rate which is the minimum displacement magnitude divided by the maximum time interval (equal to the best pick time interval plus the uncertainties at each end of the best pick interval). For example, assume the displacement magnitude is 60 to 80 km and the time interval of displacement is between 270 +/-15 Ma and 250 +/-10 Ma. The best pick displacement rate is 70km / 20Ma or 3.5 mm/yr. The minimum rate is 60km / 45Ma or 1.33 mm/yr. The uncertainty is therefore 3.5 - 1.33 = 2.17 and the displacement rate is displayed as 3.5 (+/- 2.17) mm/yr. This procedure is adopted because in some cases, as in this example, a minimum time interval (which with the maximum displacement magnitude would provide an independent maximum rate) obtained by subtracting the uncertainties from the best pick interval gives a negative time interval of displacement. In the case where displacement is constrained to begin within one time interval and end within another the uncertainties are given as the difference between the best pick rate and the minimum and maximum rates.

In almost all cases the best pick displacement rates represent minimum values because the displacement time interval represents the time within which displacement occurred. For example, a fault offsets a pluton dated at 270 Ma and is pinned by another pluton dated at 240 Ma. Displacement on the fault occurred over some time interval between 270 and 240 Ma. The true rate is therefore greater than the calculated rate which considers the entire time interval. The effect is exaggerated because displacement estimates are commonly minimum values due to the presence of cleavage and other small scale structures whose shortening effects are not considered in the displacement estimate.

The major problems in determination and interpretation of the Alleghanian displacement rate data are: 1) the use of the same time interval for displacements all along the Valley and Ridge, and 2) the relatively poorly constrained displacement magnitudes from the Piedmont province. Because displacement timing cannot be determined independently for each cross-section in the Valley and Ridge along which displacement magnitudes are determined, we must use the same timing data for each cross-section. The time interval of displacement is therefore a constant and displacement rates vary directly with displacement magnitude. Our calculated rates are therefore greater in the southern than in the central Appalachians because the displacement magnitudes are greater in the south. Rates may indeed be greater in the south but it may also be true that deformation there occurred over a longer time interval which would tend to equalize the displacement rates. In contrast to the Valley and Ridge, displacement time intervals in the Piedmont province are relatively well constrained for individual structures because of abundant metamorphic/plutonic ages but estimates of displacement magnitude are more uncertain for reasons discussed above.

Analysis of displacement and strain data

The data collected are in two forms: measurements of displacement of a given geologic feature, and measurements of finite strain of a given rock volume. The displacement measurements are generally in the form of a two-dimensional displacement vector which relates the initial location of the feature to its final location after the cessation of deformation. Measurements of finite strain usually take the form of stretches (magnitudes of the principal axes of the strain ellipse of the deformed rock volume) or ratios of stretches, and/or rotations (orientations of the strain ellipse of the deformed rock volume) of some given feature. The two different types of measurement are disposed to different methods of analysis, since they are, in a sense, two complementary halves of the data set which describes the deformation of the chosen region, in this case, the continental margin of North America.

In the case of the displacement data set, analysis of the deformation field which results from the displacements requires formulation of the linear coordinate transformation equations which describe the displacement of the data points from their initial to their final locations. The coefficients of the linear transformation equations (see equations I) are the elements of the strain matrix describing the observed displacements. These elements can be easily transformed into the same kind of strain parameters which comprise the measurements of data set two, measurements of finite strain parameters. Conversely, the analysis of the finite strain data set requires the reduction of finite strain measurements to corresponding displacements, or, in other words, exactly the reverse of the process which is applied to data set one. Once each of the two data sets has been supplied with its "complementary half," the process of incorporating the two sets into an interpretively significant regional deformation synthesis can begin in earnest.

Thus, the separate analyses of the two data sets are inverse processes. The procedure for analyzing the displacement data set, in which initial positions and displacements to final positions are known, and in which the finite strains resulting from the displacements are calculated, is known as forward modelling. The reverse process of working from final positions, represented by the finite strains of given bodies, back to the initial positions via calculation of the requisite displacement vectors, is known as inverse modelling.

We have, so far, concentrated our data analysis efforts on the methodology of forward modelling. In order to calculate strain matrices which describe the displacements of scattered points throughout an orogenic belt, it is necessary to formulate a procedure for grouping the displaced points. There are two equations which describe each linear coordinate transformation of a given point, and therefore, pairs of data are required in order to calculate exactly the coefficients of the strain matrix. If, in a given area, displacement data are available for more than two points, the system is overdetermined, and some sort of approximate solution for the strain matrix coefficients must be used. The data available for the orogenic belts which ring the North American continental margin are rather extensive, and are also characterized by complicated changes of the displacement fields throughout the Phanerozoic, and therefore, we are faced with the task of finding strain matrix coefficients for a highly overdetermined system of displacements which may or may not be the results of different deformation events. It is possible to solve for the strain matrix coefficients which best describe the displacements of all the points, assuming all the displacements can be described by just one strain matrix. Unfortunately, the results of such an exercise lead to a strain matrix which does not fit the displacement data well - in fact, if a least- squares approximating routine is used, as we have done, the strain matrix thus calculated tends to fit all the displacements equally badly. Therefore, it seems clear that it is desirable to group the displacement data into sub-sets, somehow, and then to calculate strain matrices which give valid approximations to the data contained in each of the sub-sets. In order to minimize the errors which arise through the calculation of approximate solutions to the overdetermined system, while at the same time respecting the complexity of the different deformation regimes into which any orogenic belt is divisible, we have chosen to group the displacement data into groups of three points, for which a strain matrix is calculated by means of a least- squares approximation technique. Note that this is not unlike the triangulation techniques used in analysis of geodesic measurements of presently deforming areas (Guohua and Prescott, 1986; Prescott and Yu, 1986).

A summary of the method follows:

1. Data are put in the form of initial positions of points and corresponding displacement vectors.

2. The final positions of the data points are calculated and the points are grouped into all the possible groups of three which can be formed (if N points are input, there will be (N(N - 1)(N - 2))/3! combinations of three possible). Among these trios of points, near-neighbor groupings are labelled as such.

3. For each trio of points, a strain matrix and translation vector are calculated by means of a least squares fit to the displacement vector. Thus, the initial coordinate values of the points, (x,y), are assumed to transform to the final position, (x', y'), according to the linear coordinate transformation,

$$
\begin{aligned}
x' &= ax + by + t_x \\
y' &= cx + dy + t_y
\end{aligned}
\qquad \text{(I)}
$$

and then the values of a, b, c, and d (the elements of the strain matrix), and t_x and t_y (the components of the translation vector) are calculated by minimizing the squared value of the difference between the observed (x' ,y') pairs and the calculated (x' ,y') pairs:

$$\frac{\partial}{\partial S}\left(x'_{\,obs} - x'_{\,calc}\right)^2 = 0 \qquad \text{(II)}$$
$$\frac{\partial}{\partial S}\left(y'_{\,obs} - y'_{\,calc}\right)^2 = 0$$

where S is, alternately, a, b, c, d, and t_x and t_y, and $x'_{\,calc}$ and $y'_{\,calc}$ are given by equations I.

Note that no plane strain assumption is made; therefore, area changes are permissible. Also, this procedure assumes material continuity during displacement, and homogeneity of deformation.

4. Once a strain matrix and translation vector are known for each trio of points, characteristic finite strain parameters can be calculated and graphic output, in the form of strain ellipses calculated for each trio, and placed at the centroids of the post-deformation triangles so-formed, is possible.

In order to test the accuracy and validity of the procedure, experiments were performed by constructing models of simple displacement fields which are easily interpreted. A description of one such model follows. Figure 2 is a map view of the distribution of displacements chosen for the experiment. There are seven points at which displacement data are assumed to be known (see Table 4). Basically, the displacements are those of a central rectangular area which moves laterally and undergoes a mild dextral shear relative to three fixed points. This model can be thought of as a terrane which moves and deforms relative to its surroundings, or as a thrust nappe moving within a non-deforming area.

In figure 2, the open circles represent the initial positions of the seven points, and the dotted lines between these circles indicate the near-neighbor triangle groupings. (Note that only four open circles are visible, since three of the seven points do not move and are therefore covered by solid circles.) The solid circles connected by solid lines represent the final locations of points after displacement, and the final configurations of the near-neighbor triangles.

The strain ellipses which are calculated from the three-point groupings of the seven points are shown, also in map view, in figures 3 and 4. Figure 3 displays strain ellipses for the near-neighbor combinations, while figure 4 shows ellipses for all the three-point combinations possible, including the near-neighbor groupings. As was mentioned above, the centers of the ellipses are placed at the centroids of the post-displacement triangles in both cases. The strain ellipses plotted in both figures are consistent with the geologic interpretation of the model. The topmost and bottommost ellipses are indicative of two zones of shear of opposite sense: the topmost zone is sinistral and the bottommost zone is dextral. If the motion of the rectangular area is thought to represent that of a displaced terrane, these two zones of shear would correspond to crustal-scale strike-slip zones. Thinking of the rectangle as a thrust nappe, the two shear zones can be interpreted as lateral ramps. At the two other boundaries of the moving rectangle, zones of various degrees of compression are demonstrated by the strain ellipses, as would be expected of a colliding terrane or a moving thrust sheet. The mild shear of the rectangle itself, which gives rise the the variation in the amount of compression seen along the two 'N-S' deformation belts, can also be seen in the increasing ellipticity and rotation of the more interiorly located ellipses as one moves from south to north across the displaced region.

Although the validity of the near-neighbor ellipse plot is more easily established, and despite the fact that in some cases the ellipses plotted for the all-possible trios case appear to be spurious, the two ellipse plots are highly complementary. The near-neighbor ellipses have the advantage of representing the smallest continuous triangular areas into which the displacing area can be divided, and thus they are the finest resolution division of the area. The all-possible triangles ellipses include many ellipses calculated by grouping points which are far from one another. The increased distance between the points which comprise the trio group increases the triangular area thus formed, and the probability that this area contains some structural boundary not reflected by the displacements of its vertices increases greatly. This negative feature of the all-possible trios case leads to spurious ellipses in some cases. Nevertheless, as figures 3 and 4 amply demonstrate, the greater number of ellipses in figure 4 aids greatly in the interpretation of figure 3, while figure 3 makes clear the spurious ellipses of figure 4.

Beyond the ellipse plots, we have created contoured plots of the various strain parameters (principal stretches, ellipticity, dilatation, and rotation). Figures 5 and 6 show two such plots, both contours of ellipticity. Figure 5 represents ellipticities arising from near-neighbor trio groupings, while figure 6 is of ellipticities calculated from all-possible combinations of three groupings. Although these plots are more difficult to interpret than the ellipse plots, they do give some indication of the two zones of lateral shear described above, and of the dextral shear of the interior rectangle. The contour plots may also aid in the discovery of zones of previously unrecognized displacement.

REFERENCES CITED

Allmendinger, R. W., 1982, Analysis of microstructures in the Meade plate of the Idaho-Wyoming foreland thrust belt, USA.: Tectonophysics, v. 85, p. 221-251.

Allmendinger, R. W., Farmer, H., Hauser, E., Sharp, J., Von Tish, D., Oliver, J., and Kaufman, S., 1986, Phanerozoic tectonics of the Basin and Range-Colorado Plateau transition from COCORP data and geologic data: a review., Reflection Siesmology: The Continental Crust Geodynamics Series Vol. 14., American Geophysical Union, p. 257-267.

Allmendinger, R. W., Miller, D. M., and Jordan, T. E., 1985, Known and inferred Mesozoic deformation in the hinterland of the Sevier Belt, Noerthwest Utah., *in* Geology of Northwest Utah, Southern Idaho, and Northeast Nevada, Kerns, G. J., ed., Utah Geological Association, Pub. no. 13, p. 21-34.

Anderson, F.D., 1972, The Catamaran fault, north-central New Brunswick: Canadian Journal of Earth Science, v. 9, p. 1278-1286.

Archibald, D. A., Glover, J. K., Price, R. A., Farrar, E., and Carmichael, D. M., 1983, Geochronology and tectonic implications of magmatism and metamorphism, southern Kootenay arc and neighboring regions, southeastern British Columbia. Part 1: Jurassic to mid-Cretaceous.: Canadian Journal of Earth Science, v. 20, p. 1891-1913.

Armstrong, R. L., 1975, Precambrian (1500my old) rocks of central Idaho - the Salmon River Arch and its role in Cordilleran sedimentation and tectonics.: American Journal of Science, v. 275-A, p. 437-467.

Armstrong, R. L., Taubeneck, W. H., and Hales, P. O., 1977, Rb-Sr and K-Ar geochronometry of Mesozoic granitic rocks and their Sr isotopic composition, Oregon, Washington, and Idaho.: Geological Society of America Bulletin, v. 88, p. 321-331.

Aronson, J., and Elliott, W., 1985, Dating of thrusting in the Idaho-Wyoming belt: K-Ar ages of authigenic illite formed by burial beneath thrust sheets and or synorogenic sediments.: Geological Society of America Abstracts with Programs, v. 17, p. 207.

Bally, A. W., 1984, Tectogenese et sismique reflexion: Bulletin Societe Geologique de France, v. 26, p. 279-285.

Bartholomew, M.J., 1987, Structural evolution of the Pulaski thrust system, southwestern Virginia:

Geological Society of America Bulletin, v. 99, p. 491-510.

Boyer, S.E., and Elliot, D., 1982, Thrust systems: American Association of Petroleum Geologists Bulletin, v. 66, p. 1196-1230.

Bregman, M. L., 1976, Change in tectonic style along the Montana thrust belt: Geology, v. 4, p. 775-778.

Brown, R. L., Journeay, M., Lane, L. S., Murphy, D. C., and Rees, C. J., 1986, Obduction, backfolding, and piggyback thrusting in the metamorphic hinterland of the southeastern Canadian Cordillera.: Journal of Structural Geology, v. 8, p. 255-268.

Campbell, R. B., Mountjoy, E. W., and Struik, L. C., 1982, Structural cross-section through the south-central Rocky and Cariboo Mountains: Geological Survey of Canada Open File Report 844.

Chase, R. B., Bickford, M. E., and Arruda, E. C., 1983, Tectonic implications of Tertiary intrusion and shearing within the Bitterroot Dome, northeastern Idaho Batholith.: Journal of Geology, v. 91, p. 462-470.

Chesson, S. A., Billodeau, W. L., and Wells, J. D., 1984, Cambrian stratigraphy of the Libby and Alberton areas, northwestern Montana., *in* Northwest Montana and Adjacent Canada, McBane, J. D., and Garrison, P. B., eds., Montana Geological Society 1984 Field Conference and Symposium, p. 121-124.

Dallmeyer, R. D., Snoke, A. W., and McKee, E. H., 1986, The Mesozoic-Cenozoic tectonothermal evolution of the Ruby Mountains, East Humboldt Range, Nevada: A cordilleran metamorphic core complex.: Tectonics, v. 5, p. 931-954.

DeCelles, P. G., 1986, Sedimentation in a tectonically partitioned, non-marine foreland basin: Lower Cretaceous Kootenai Formation, southwestern Montana.: Geological Society of America Bulletin, v. 97, p. 911-931.

DeCelles, P. G., Tolson, R. B., Graham, S. A., Smith, G. A., Ingersoll, R. V., White, J., Schmidt, C. J., Rice, R., Moxon, I., Lemke, L., Handschy, J. W., Follo, M. F., Edwards, D. P., Cavazza, W., Caldwell, M., and Bargar, E., 1987, Laramide thrust-generated alluvial-fan sedimentation, Sphinx Conglomerate, southwestern Montana.: American Association of Petroleum Geologists Bulletin, v. 71, p. 135-155.

Dennis, A.J., and Secor, D.T., 1985, Genesis and geometry of an Alleghanian shear zone in the central South Carolina Piedmont: Geological Society of America Abstracts with Programs, v. 17, p. 87.

Dixon, J. S., 1982, Regional structural synthesis, Wyoming salient of western overthrust belt.: American Association of Petroleum Geologists Bulletin, v. 66, p. 1560-1580.

Druhan, R.M., Butler, J.R., Horton, J.W., and Stoddard, E.F., 1988, Tectonic significance of the Nutbush Creek fault zone, eastern Piedmont of North Carolina and Virginia: Geological Society of America Abstracts with Programs, v. 20, p. 261.

Fountain, D. M., and McDonough, D., 1984, Crustal structure of northwestern Montana and southern Alberta and British Columbia based on interpretation of gravity data., *in* Northwest Montana and Adjacent Canada, McBane, J. D., and Garrison, P. B., eds., Montana Geological Society 1984 Field Conference and Symposium, p. 253-258.

Fox, K. F., Jr., Rinehart, C. D., and Engels, J. C., 1977, Plutonism and orogeny in north-central Washington - timing and regional context.: United States Geological Survey Professional Paper 989, 27 p.

Gates, A.E., 1987, Transpressional dome formation in the southwest Virginia Piedmont: American Journal of Science, v. 287, p. 927-949.

Gates, A.E., Simpson, C., and Glover, L., 1986, Appalachian Carboniferous strike-slip faults: an example from Brookneal, Virginia: Tectonics, v. 5, p. 119-133.

Geiger, B. C., 1985, Development of cleavage in the frontal fold and thrust belt, southwestern Montana: evidence for a foreland buttress.: Geological Society of America Abstracts with Programs, v. 17, p. 220.

Glover, L., III, and Gates, A.E., 1987, Alleghanian orogeny in the central and southern Appalachians: Geological Society of America Abstracts with Programs, v. 19, p. 86.

Gockley, C., 1985, Fold geometry and strain results: Big Elk Mountain anticline, Absaroka thrust sheet (SE Idaho).: Geological Society of America Abstracts with Programs, v. 17, p. 220.

Goldburg, B. L., 1984, Displacement transfer between thrust faults near the Sun River in the Sawtooth Range, northwestern Montana., *in* Northwest Montana and Adjacent Canada, McBane, J. D., and Garrison, P. B., eds., Montana Geological Society 1984 Field Conference and Symposium, p. 211-

220.

Guohua, G., and Prescott, W., 1986, Discussion on displacement analysis: detection of crustal deformation: Journal of Geophysical Research, v. 91, p. 7439-7447.

Gwinn, V.E., 1970, Kinematic patterns and estimates of lateral shortening, Valley and Ridge and Great Valley provinces, central Appalachians, south-central Pennsylvania, *in* Studies of Appalachian Geology: Central and Southern, Fisher, G.W., Pettijohn, F.J., Reed, J.C., Jr., and Weaver, K.N., eds., Interscience Publishers, New York, p. 127-147.

Harrison, J. E., Kleinkopf, M. D., and Wells, J. D., 1980, Phanerozoic thrusting in Proterozoic Belt rocks, northwestern United States.: Geology, v. 8, p. 407-411.

Heller, P. L., Bowdler, S. S., Chambers, H. P., Coogan, J. C., Hogen, E. S., Shuster, M. W., Winslow, N. S., and Lawton, T. F., 1986, Time of initial thrusting in the Sevier orogenic belt, Idaho-Wyoming and Utah.: Geology, v. 14, p. 388-391.

Hoffman, J., Hower, J., and Aronson, J. L., 1976, Radiometric dating of time of thrusting in the disturbed belt of Montana.: Geology, v. 4, p. 16-20.

Hyndman, D. W., 1979, Major tectonic elements and tectonic problems along the line of section from northeastern Oregon to west-central Montana.: Geological Society of America Map and Chart Series, Map MC-28C, 11 p.

Johnson, C., 1985, Internal deformation and finite strain estimates of an exposed frontal thrust fold in the overthrust belt.: Geological Society of America Abstracts with Programs, v. 17, p. 227.

Jordan, T. E., 1981, Thrust loads and foreland basin evolution, Cretaceous, western United States.: American Association of Petroleum Geologists Bulletin, v. 65, p. 2506-2520.

Journeay, M., and Brown, R. L., 1986, Major tectonic boundaries of the Omineca belt in southern British Columbia: a progress report.: Current Research Part A, Geological Survey of Canada Paper 86-1A, p. 81-88.

Kulander, B.R., and Dean, S.L., 1986, Structure and tectonics of central and southern Appalachian Valley and Ridge and Plateau provinces: American Association of Petroleum Geologists Bulletin, v. 70, p. 1674-1684.

Lageson, D. R., Schmidt, C. J., Dresser, H. W., Welker, M., Berg, R. B., and James, H. L., 1983a, Road Log 1: Bozeman to Helena via Battle Ridge Pass, White Sulfer Springs, Townsend and Toston., Guidebook of the fold and thrust belt, west-central Montana, compiled by D. L. Smith, Montana Bureau of Mines and Geology Special Publication 86, p. 1-31.

—— 1983b, Road Log 2: Helena to Bozeman via MacDonald Pass, Garrison, Deer Lodge, Butte, Homestake Pass, Jefferson Canyon, and Three Forks., Guidebook of the fold and thrust belt, west-central Montana, compiled by D. L. Smith, Montana Bureau of Mines and Geology Special Publication 86, p. 32-82.

Lapham, D.M., and Bassett, W.A., 1964, K-Ar dating of rocks and tectonic events in the Piedmont of southeastern Pennsylvania: Geological Society of America Bulletin, v. 75, p. 661-668.

Miller, E. L., Gans, P.B., Wright, J. E., and Sutter, J. F., in press, 1988, Metamorphic history of the east-central Basin and Range province: tectonic setting and relationship to magmatism., *in* Rubey Volume 7, Metamorphism and Crustal Evolution of the Western United States., Ernst, W. G., ed., University of California, Los Angeles, Calif.

Miller, F. K., and Engels, J. C., 1975, Distribution and trends of discordant ages of the plutonic rocks of northeastern Washington and northern Idaho.: Geological Society of America Bulletin, v. 86, p. 517-528.

Mitra, G., 1979, Ductile deformation zones in Blue Ridge basement rocks and estimation of finite strain: GSAB, Part I, v. 90, p. 935-951.

Mitra, S., 1988, Three-dimensional geometry and kinematic evolution of the Pine Mountain thrust system, southern Appalachians: Geological Society of America Bulletin, v. 100, p. 72-95.

Mudge, M. R., and Earhart, R. L., 1980, The Lewis thrust fault in the disturbed belt, northwestern Montana.: United States Geological Survey Professional Paper 1174, 18 p.

Nichols, D. J., Perry, W. J. JR., and Haley, J. C., 1985, Reinterpretation of the palynology and age of Laramide syntectonic depsits, southwestern Montana, and revision of the Beaverhead Group.: Geology, v. 13, p. 149-153.

O'Hara, K., and Gromet, L.P., 1985, Two distinct late Precambrian (Avalonian) terranes in southeastern

New England and their Late Paleozoic juxtaposition: American Journal of Science, v. 285, p. 673-709.

Palmer, A.R., 1983, The decade of North American geology 1983 time scale: Geology, v. 11, p. 503-504.

Parkinson, D., 1985, Geochronology of the western side of the Okanogan metamorphic core complex, southern British Columbia.: Geological Society of America Abstracts with Programs, v. 17, p. 399.

Parrish, R. P., and Wheeler, J. O., 1983, A U-Pb zircon age from the Kuskanax batholith, southeastern British Columbia.: Canadian Journal of Earth Science, v. 20, p. 1751-1756.

Perry, W. J., Jr., Wardlaw, B. R., Bostick, N. H., and Maughan, E. K., 1983, Structure, burial history, and petroleum potential of frontal thrust belt and adjacent foreland, southwest Montana.: American Association of Petroleum Geologists Bulletin, v. 67, p. 725-743.

Prescott, W., and Yu, S., 1986, Geodetic measurements of horizontal deformation in the northern San Francisco Bay region, California: Journal of Geophysical Research, v. 91, p. 7475-7485.

Price, R. A., 1981, The Canadian foreland thrust and fold belt in the southern Canadian Rocky Mountains., *in* Thrust and Nappe Tectonics, McClay, K. R., and Price, N. J., eds., Geological Society of London Special Publication 9, p. 427-448.

Price, R. A., and Mountjoy, E. W., 1970, Geologic structure of the Canadian Rocky Mountains between Bow and Athabasca rivers - a progress report., *in* Structure of the southern Canadian cordillera, Wheeler, J. O., ed., Geological Association of Canada Special Paper 6, p. 7-25.

Royse, F., Jr., Warner, M. A., and Reese, D. L., 1975, Thrust belt structural geometryy and related stratigraphic problems, Wyoming-Idaho-northern Utah., *in* Symposium on deep drilling frontiers in the central Rocky Mountains, Bolyard, D. W., ed., Rocky Mountain Association of Geologists, p. 41-55.

Russell, G.S., Russell, C.W., and Farrar, S.S., 1985, Alleghanian deformation and metamorphism in the eastern North Carolina Piedmont: Geological Society of America Bulletin, v. 96, p. 381-387.

Schmidt, C. J., James, C., and Shearer, J. N., 1981, Estimate of displacement in major zone of tear-thrusting in fold and thrust belt, southwest Montana.: American Association of Petroleum Geologists Bulletin, v. 65, p. 986-987.

Secor, D.T., Snoke, A.W., Bramlett, K.W., Costello, O.P., and Kimbrell, O.P., 1986b, Character of the

Alleghanian orogeny in the southern Appalachians: Part I. Alleghanian deformation in the eastern Piedmont of South Carolina: Geological Society of America Bulletin, v. 97, p. 1319-1328.

Snoke, A. W., and Miller, D. M., in press, 1988, Metamorphic and tectonic history of the northeastern Great Basin., *in* Rubey Vol. 7, Metamorphism and Crustal Evolution of the Western United States., Ernst, G., ed., University of California, Los Angeles, Calif.

Speed, R. C., Elison, M. W., and Heck, F. R., in press, 1988, Phanerozoic tectonic evolution of the Great Basin., *in* Rubey Volume 7, Metamorphism and Crustal Evolution of the Western United States, Ernst, G., ed., University of California, Los Angeles, Calif.

Speed, R. C., and Sleep, N. H., 1982, Antler orogeny and foreland basin: A model.: Geological Society of America Bulletin, v. 93, p. 815-828.

Thompson, R. I., 1979, A structural interpretation across part of the northern Rocky Mountains, British Columbia, Canada.: Canadian Journal of Earth Science, v. 16, p. 1228-1241.

Valentino, D.W., 1988, Rosemont shear zone, central Appalachian Piedmont: dextral displacement of the Willmington Complex: Geological Society of America Abstracts with Programs, v. 20, p. 77.

Villien, A., and Kligfield, R. M., 1986, Thrusting and synorogenic sedimentation in central Utah., *in* Paleotectonics and sedimentation in the Rocky Mountain Region, United States, Peterson, J. A., ed., American Association of Petroleum Geologists Memoir 41, p. 281-307.

Webb, G.W., 1969, Paleozoic wrench faults in Canadian Appalachians, *in* North Atlantic - Geology and Continental Drift, Kay, M., ed., American Association of Petroleum Geologists Memoir 12, p. 754-786.

Wetherill, G.W., Tilton, G.R., Davis, G.L., Hart, S.R., and Hopson, C.A., 1966, Age measurements in the Maryland Piedmont: Journal of Geophysical Research, v. 71, p. 2139-2155.

Wiltschko, D. V., and Dorr, J. A., Jr., 1983, Timing of deformation in overthrust belt and foreland of Idaho, Wyoming and Utah.: American Association of Petroleum Geologists Bulletin, v. 67, p. 1304-1322.

Woodward, N. B., 1986, Thrust fault geometry of the Snake River Range, Idaho and Wyoming.: Geological Society of America Bulletin, v. 97, p. 178-193.

Yeo, G.M., and Gao Ruixing, 1986, Late Carboniferous dextral movement on the Cobequid-Hollow fault system, Nova Scotia: evidence and implications: Current Research, Part A, Geological Survey of Canada, Paper 86-1A, p. 399-410.

TABLE 1. Displacement magnitude, period and rate for various positions along the fold and thrust belt.

Location (see Fig. 1)	Horizontal Displacement	Reference	Displacement Onset	Reference	Displacement End	Reference	Displacement period	Displacement rate
	$120km_a$	Brown et al, 1986; Campbell et al, 1982	$152\pm7_1$	Thompson, 1979	66_2	Thompson, 1979	86my	$1.40\pm.37$mm/yr
B	$200km_a$	Price, 1981; Price and Mountjoy, 1970; Brown et al, 1986	$152\pm7_1$	Price, 1981	$40\pm2_3$ $58\pm2_2$	Price,1981 "	112my	$2.13\pm.81$mm/yr
C	$170\text{-}180km_a$	Harrison et al, 1980; Bally, 1984	$115\pm3_1$ $154\pm2_4$	DeCelles,1986 "	$49\pm9_3$ 56_5 46.3_6 58.3_6	Bally, 1984 Hoffman et al, 1976 Mudge and Earhart, 1980	66my	2.65 ± 1.18mm/yr
D	$150\text{-}180_{b,c}$	Chesson et al, 1984; Hyndman, 1979; Chase et al, 1983; Lagenson et al, 1984a,b	$115\pm3_1$ $154\pm2_4$	DeCelles, 1986 "	56_2 $57.7\pm1.4_6$	DeCelles et al, 1987 "	59my	2.8 ± 1.3mm/yr
E	$140\text{-}150km_a$	Jordon, 1981; Royse et al, 1975	$145\pm5_1$	Armstrong and Cressman, 1983	$52\pm2_3$	Wiltschko and Dorr, 1983; Royse et al, 1975	93my	$1.56\pm.12$mm/yr
			$*116\pm3$	Heller et al, 1986			62my	$2.34\pm.94$mm/yr
F	$100\text{-}120km_a$	Allmendinger et al, 1986	$105\pm7_1$	Villien and Kligfield, 1986	$55\pm3_3$	Villien and Kligfield,1986	50my	$2.2\pm.56$mm/yr

TABLE 1. (cont)

a) displacement magnitude by palinspastic reconstruction

b) displacement magnitude by summation of individual thrust displacements

c) displacement magnitude by correlation to north and/or south

1) displacement onset by earliest foreland basin sediments

2) displacement end by cessation of foreland basin sedimentation

3) displacement end by overlapping strata

4) possible displacement onset by youngest strata beneath foreland basin sediments

5) displacement end by youngest burial metamorphism - radiometric age

6) displacement end by date of cross-cutting intrusion - radiometric age

TABLE 2. Events and timing west of the fold and thrust belt.

Location	Event	Timing	Reference
west of lines B and C	syn-metamorphic plutonism	175-165my	Archibald et al, 1983
	burial metamorphic and uplift	166-156my	"
	syn-metamorphic deformation	160-150my	Journeay and Brown, 1986
	Kuskanax batholith (syn-deformational)	173my	Parrish and Wheeler, 1983
	syn-deformational plutonism	170-161my	Parkinson, 1985
	major plutonic and metamorphic pulse	110-94my	Fox et al, 1977
	post-deformational plutonism	48-49my	"
	major plutonic pulse	101-93my	Miller and Engels, 1975
	major plutonic pulse	51-47my	"
west of line D	metamorphism in Salmon River Arch	115-103my	Armstrong,1975
west of lines E and F	metamorphism in Ruby Range	160my	Dallmeyer et al, 1986
	plutonism, deformation and metamorphism in eastern Great Basin	170my onset	Allmendinger et al, 1985 Miller et al, 1988 Snoke and Miller, 1988
	plutonism and deformation in north-central Nevada	170-150my	Speed et al, 1988

- onset of metamorphism, plutonism, and deformation was middle to late Jurassic in southern British Columbia, northern Washington and Idaho, and western Utah and eastern Nevada.
- onset of metamorphism and deformation in central Idaho is only constrained to be pre-late Cretaceous.

Table 3: Alleghanian displacement rates

	Name of structure	Displacement direction	Displacement magnitude min./average/max. (km)	Displacement timing beginning/end (Ma)	Disp. time interval min./pick/max. (Ma)	Displacement rate (mm/yr)
orogen-parallel displacements	Catamaran fault [1]	S73W	7 / 12 / 16	374±18 / 360±10 [A]	--/ 14 / 42 [a]	.82 ±.65
	Bellisle fault [2]	S44W	32 / 48 / 64	355 ±9 / 345 ±15 [A]	--/ 10 / 34	4.80 ±4.03
	Peekaboo-Berry Mills fault [2]	S44W	32 / 35 / 38	355 ±9 / 345 ±15 [A]	--/ 10 / 34	3.50 ±2.56
	Clover Hill fault [2]	S57W	9 / 12 / 15 [b]	355 ±9 / 345 ±15 [A]	--/ 10 / 34	1.20 ±.94
	Cobequid-Chedabucto fault [2]	west	157 / 210 / 263 [b]	374 ±18 / 296 ±10 [A,B]	--/ 78 / 106	2.69 ±1.21
	Harvey-Hopewell fault[2]					
	dextral displacement	S42W	60 / 80 / 100 [b]	350 ±10 / 320 ±20 [A]	--/ 30 / 60	2.67 ±1.67
	sinistral disp.	N42E	greater than 16	310 ±15 / 300 ±12 [A]	--/ 10 / 37	>1.60 ±1.17
	Long Range fault[2]	S51W	105 / 181 / 257	333 ±22 / 300 ±12 [A]	--/ 33 / 67	5.48 ±3.92
	Hollow fault[3]	S75W	26 / 35 / 44 [b]	345 ±15 / 298 ±12 [A]	--/ 47 / 74	.74 ±.39
	Hope Valley shear zone[4]	south	25 / 45 / 65	290 / 276 [A,B]	--/ 14 / 21[c]	3.21 ±2.02
	Rosemont shear zone	S45W	greater than 25[5]	330 ±16 / 290[6,C,d]	--/ 40 / 56	>.63 ±.18
	Baltimore gneiss domes	S65W	112 / 150 / 188[7,b]	310 ±25 / 280[8,C,d]	--/ 30 / 55	5.00 ±2.96
	Brookneal shear zone[9]	S50W	greater than 17	324 ±3 / 300 ±5[B,C]	--/ 24 / 32	>.71 ±.18
	Nutbush Creek shear zone	S20W	120 / 160 / 200[10,b]	313-285 / 251-238[11,B,C]	34 / 54 / 75	2.96 $^{+2.92}_{-1.36}$
	Modoc zone	S67W	18 / 25 / 32 [12,b]	292 ±15 / 268 ±5[13,B,C]	--/ 24 / 44	1.04 ±.63
orogen-normal displacement	Valley and Ridge					
	section 1 [14]	N67W	103 / 138 / 173[b,e]	303 ±13 / 265-235[f]	--/ 38 / 81	3.63 ±2.36
	section 2 [15,16]	N50W	150 / 208 / 267[e]	303 ±13 / 265-235	--/ 38 / 81	5.47 ±3.55
	section 3 [15,16]	N28W	213 / 259 / 305[e]	303 ±13 / 265-235	--/ 38 / 81	6.82 ±4.03
	section 4[15,17,18]	N35W	202 / 269 / 336 [b,e]	303 ±13 / 265-235	--/ 38 / 81	7.08 ±4.59

1. Anderson, 1972
2. Webb, 1969
3. Yeo and Gao Ruixing, 1986
4. O'Hara and Gromet, 1985
5. Valentino, 1988
6. Lapham and Bassett, 1964
7. Glover and Gates, 1987
8. Wetherill and others, 1966
9. Gates and others, 1986
10. Druhan and others, 1988
11. Russell and others, 1985
12. Dennis and Secor, 1985
13. Secor and others, 1986b
14. Gwinn, 1970; Mitra, 1979
15. Kulander and Dean, 1986
16. Bartholomew, 1987
17. Boyer and Elliot, 1982
18. Mitra, 1988

Structures and cross-section lines are located on Fig. 1. Strike-slip displacement directions are dextral unless noted otherwise and give the transport direction with respect to North America of rocks outboard the fault.

A. stratal timing constraints
B. intrusive ages
C. metmorphic ages

a. minimum is given only when the displacement begins within one time interval and ends within another (see text)
b. min. and max. are 25% more and 25% less than a single value given in the cited reference
c. max. is best pick interval plus 50%
d. deformation assumed to occur with last major thermal event
e. magnitudes are the sum of displacements across discrete parts of the Valley and Ridge
f. see text for explanation of this timing

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Table 4		
Initial position	Final position	Displacement
(x,y)	(x',y')	(u,v)
220.0,350.0	220.0,350.0	0.0.0.0
200.0,300.0	230.0,300.0	30.0,0.0
275.0,300.0	300.0,300.0	25.0
375.0,275.0	375.0,275.0	0.0,0.0
201.0,249.0	223.0,249.0	22.0,0.0
274.0,250.0	292.0,250.0	18.0,0.0
221.0,198.0	221.0,198.0	0.0,0.0

Figure Captions

Figure 1. Map of the western North American Cordillera showing structural and thermal features discussed in text. Lines with corresponding letters A through F are the locations for which displacement magnitude, period, and rate data are presented.

Figure 2: 2a shows the physiographic provinces of the southern and central Appalachians refered to in text. 2b shows the various tectonic features referred to in the text and Table 1: 1. Long Range-Cabot fault, 2. Catamaran fault, 3. Bellisle fault, 4. Peekaboo-Berry Mills fault, 5. Clover Hill fault, 6. Harvey-Hopewell fault, 7. Hollow fault, 8. Cobequid-Chedabucto fault, 9. St. John fold and thrust belt, 10. Norumbega fault, 11. Kingman fault, 12. Kearsarge-central Maine synclinorium, 13. Bloody Bluff fault, 14. Lake Char / Clinton-Newbury fault zone, 15. Honey Hill fault, 16. Hope Valley shear zone, 17. Narragansett Basin, 18. Martic zone, 19. Rosemont shear zone, 20. Baltimore gneiss domes, 21. Hylas shear zone, 22. Brookneal shear zone, 23. Bowens Creek shear zone, 24. Hollister shear zone, 25. Nutbush Creek shear zone, 26. Modoc zone, 27. Kings Mountain Belt shear zone, 28. Brevard zone, 29. Goats Rock shear zone, 30. Towaliga shear zone, 31. Roanoke Recess, 32. Blue Ridge thrust, 33. Pulaski thrust, 34. Grandfather Mountain Window, 35. Pine Mountain thrust, 36. western limit of Alleghanian deformation. Circled numbers are cross-section lines from Table 3. 2c shows the method of determining the motion with respect to undeformed North America of a point (5) on the outboard side of the orogen that is displaced along an orogen-parallel strike-slip fault.

Figure 3: Map view of the configurations of initial and final triangles formed from near-neighbor grouping of the initial displacement data. The initial locations of the points are the hollow circles, and the final positions are the filled circles. The dotted lines are drawn between the initial triangles, and the solid lines between the final triangles. The scale on the axes is distance from the arbitrarily chosen origin. The y-axis is parallel to North. Note that three of the points do not move, and therefore three hollow circles are covered by solid circles.

Figure 4: Map view of the strain ellipses calculated from the near-neighbor three point groupings. Scale and axes as in figure 2.

Figure 5: Map view of the strain ellipses calculated from all the possible three point groupings of the data.

Figure 6: Contoured map view of ellipticities of strain ellipses calculated from the near-neighbor triangles case.

Figure 7: Contoured map view of ellipticities of strain ellipses calculated from all the possible three point groupings of the data.

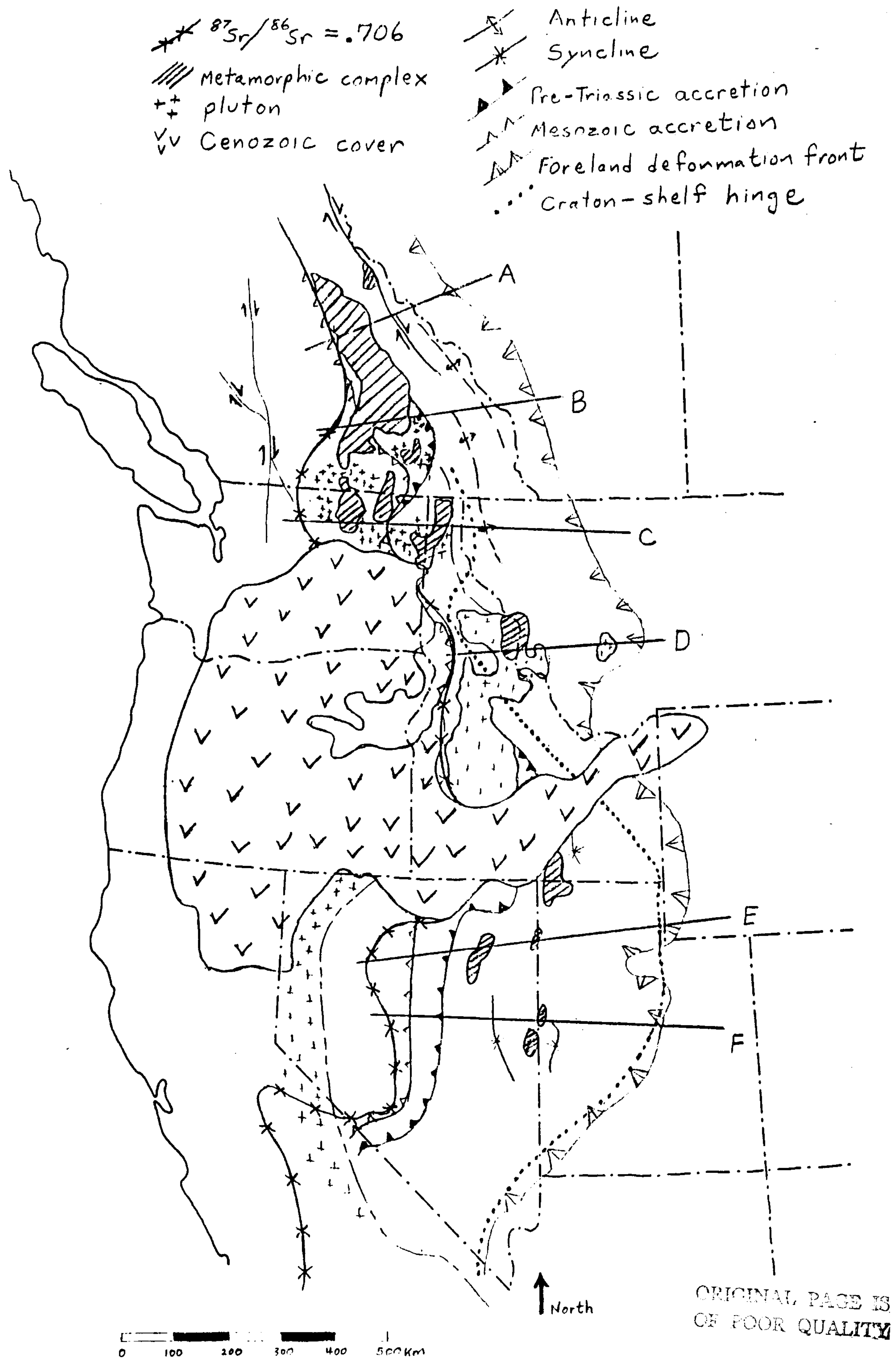
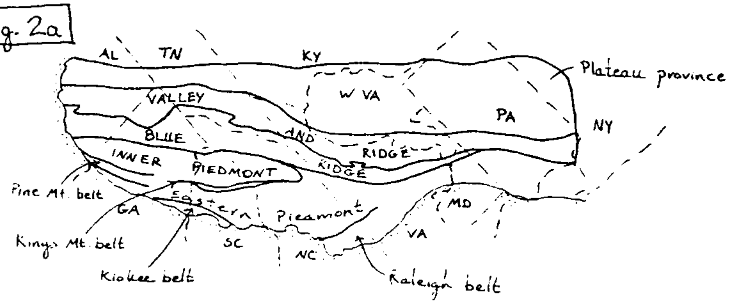

$^{87}Sr/^{86}Sr = .706$
Metamorphic complex
pluton
Cenozoic cover
Anticline
Syncline
Pre-Triassic accretion
Mesozoic accretion
Foreland defonmation front
Craton-shelf hinge
A
B
C
D
E
F
North
0 100 200 300 400 500 Km

FIG 1

Fig. 2a



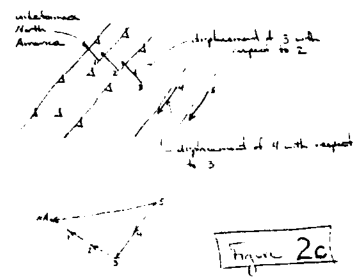
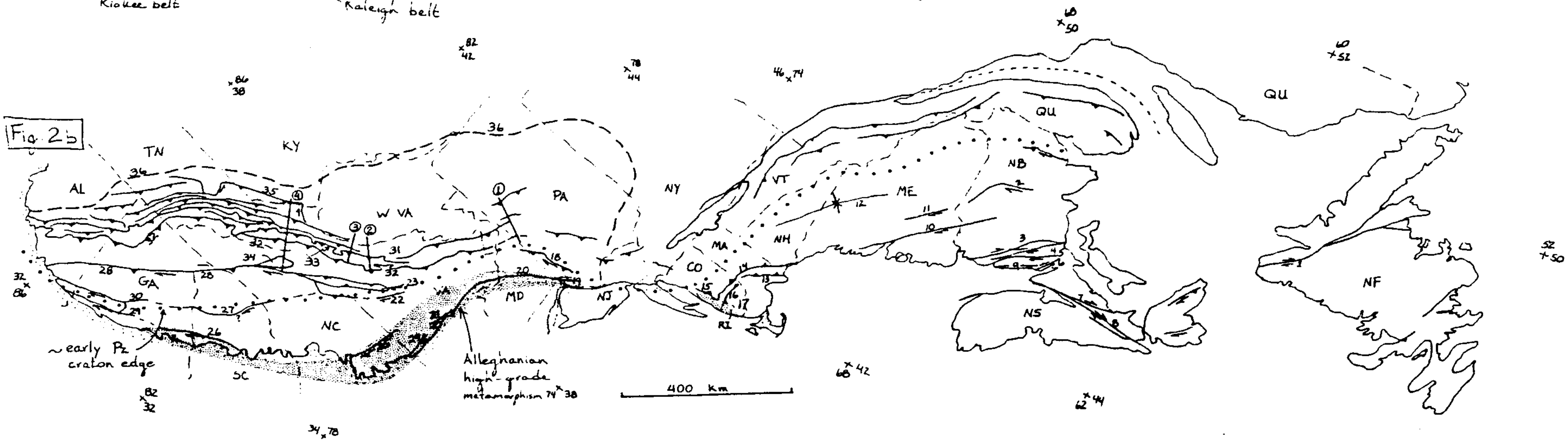


Figure 2c

Fig. 2b



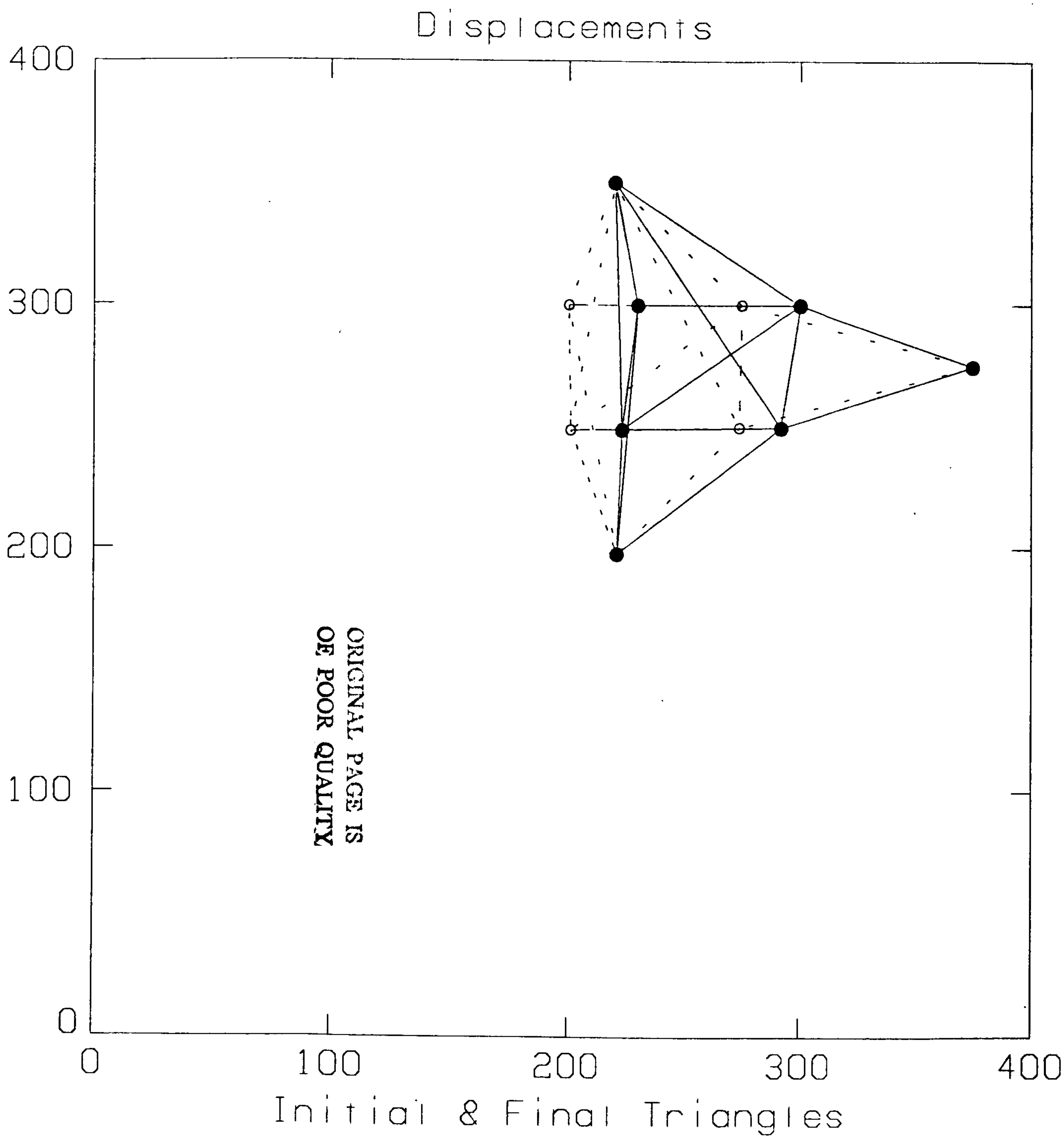


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Fig. 3

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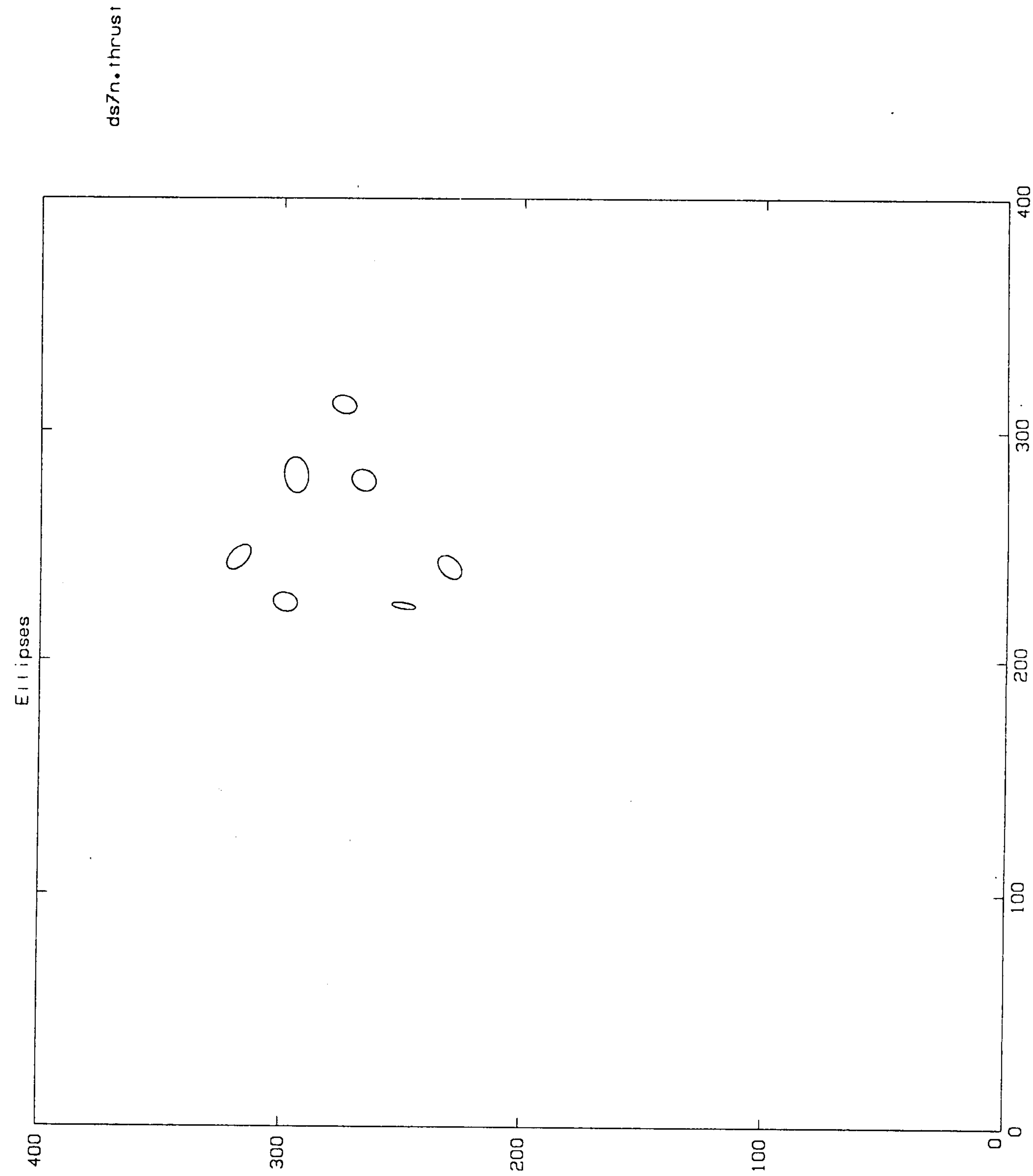
Fig 4



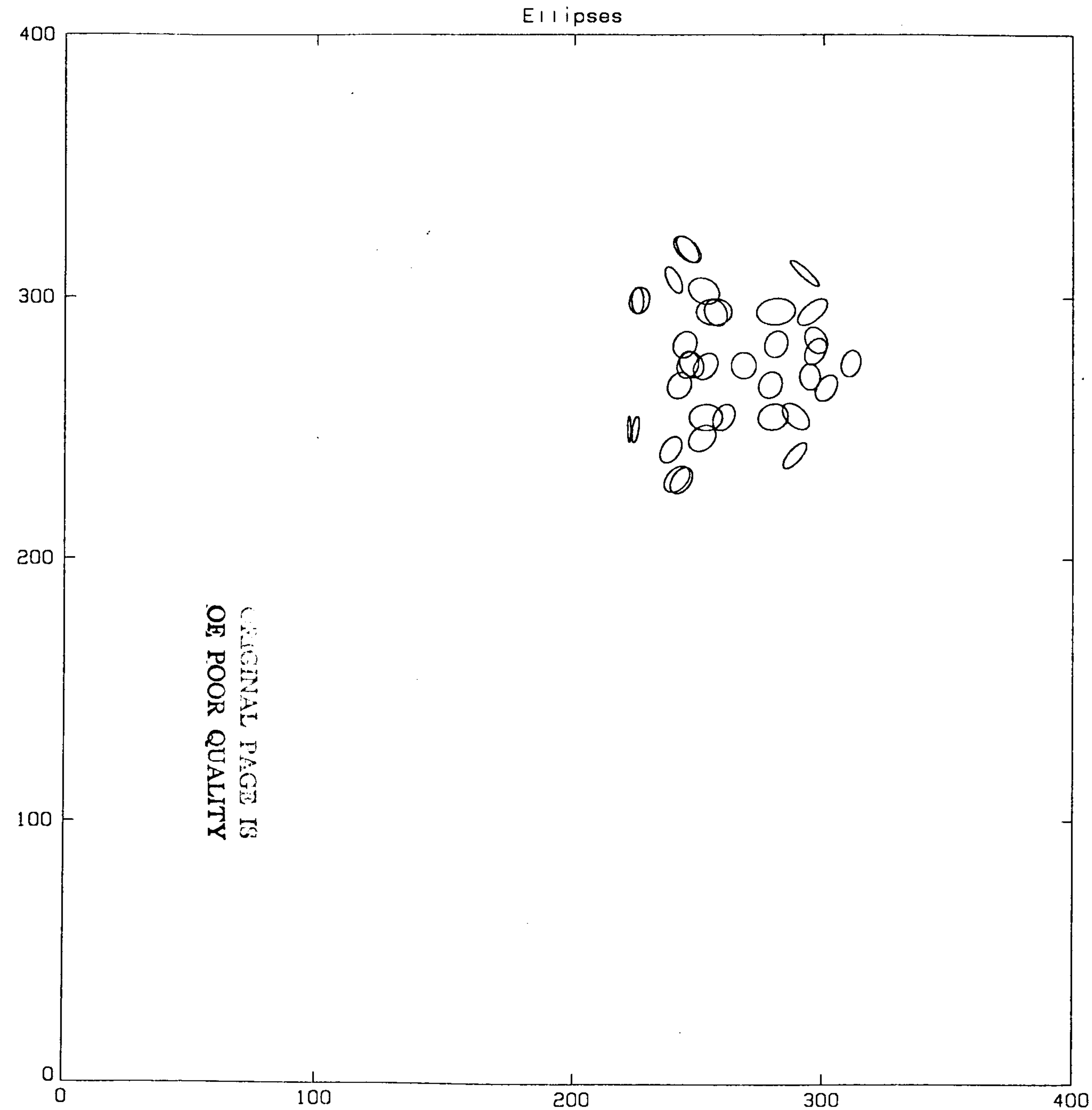


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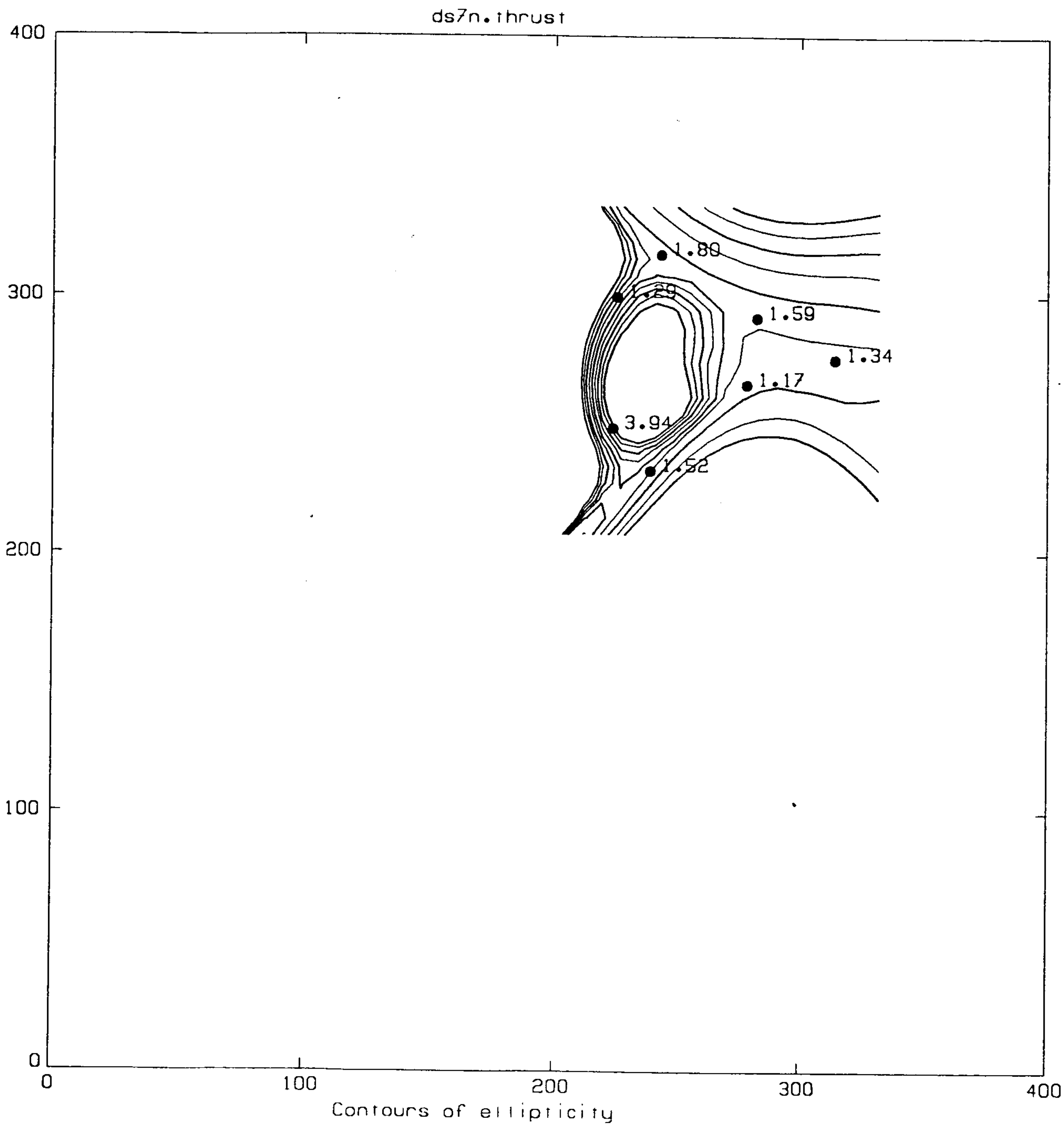
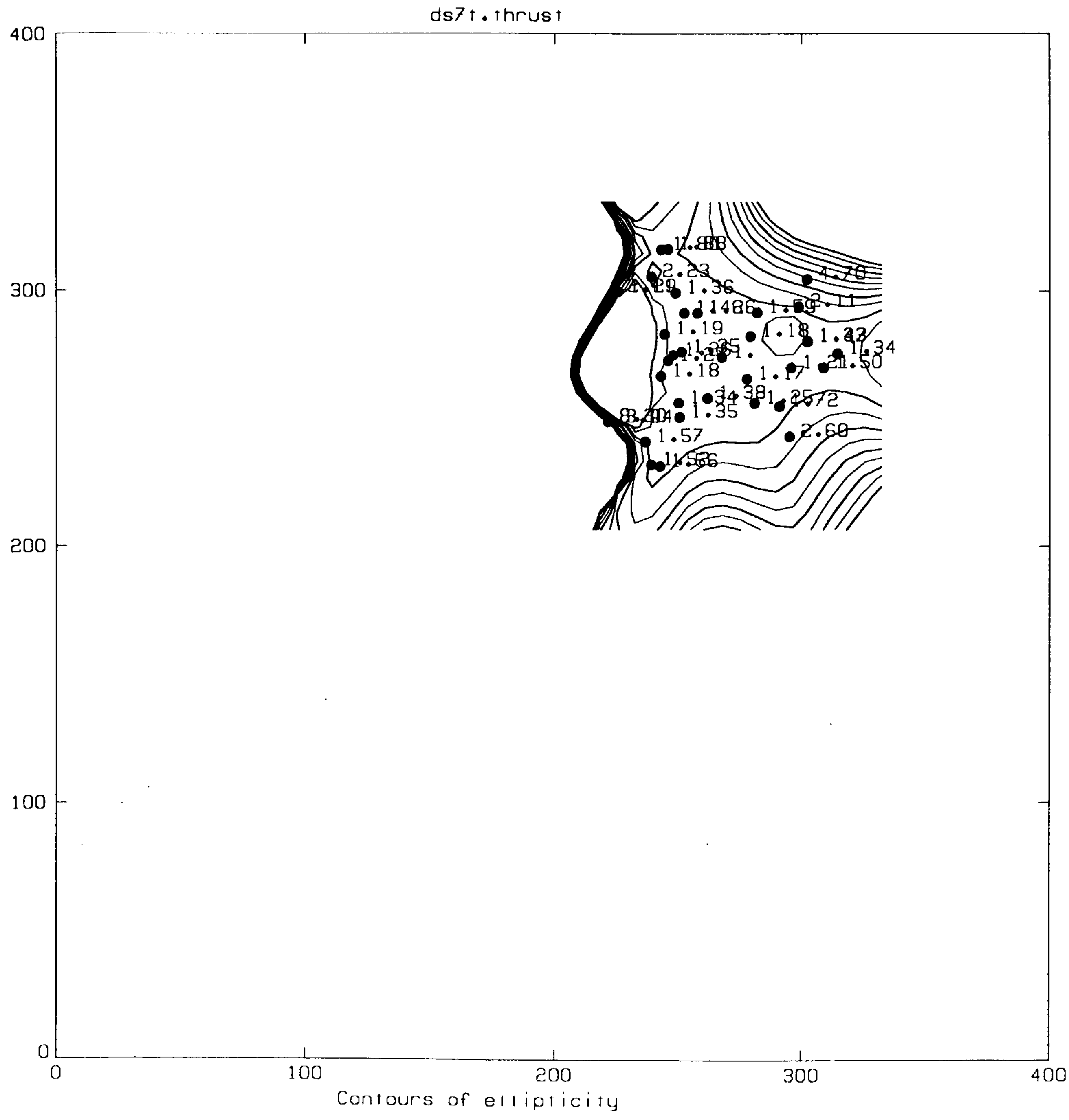
Fig 5



Ellipticity

Fig 6



Ellipticity

Fig 7

Appendix 2

DISPLACEMENT DISTRIBUTION IN THE CORDILLERA OF WESTERN NORTH AMERICA

M.W.Elison
Department of Geological Sciences
Northwestern University
Evanston Il. 60201

Introduction

This paper presents an analysis of the timing, amount and rate of continental margin deformation, exclusive of accreted terranes, and of the horizontal and vertical distribution of displacements within the continental margin. Possible models of the relationship between deformation recorded in the foreland fold and thrust belt and deformation recorded in areas underlain by sialic crust to the west of the foreland are considered.

Studies along the western North American Cordillera from Nevada to southern British Columbia demonstrate striking structural similarity, not only within the foreland fold and thrust belt, but also in areas to the west. Examination of foreland shortening for six positions along the Cordillera indicate that displacement magnitude decreases regularly from north to south. Concomitant with the decrease in displacement magnitude, the width of the area between the eastern limit of accreted terranes and the craton-shelf hinge increases. The north to south increase in width of this area is consistent when the effects of Tertiary extension are removed.

The western part of the continental margin, east of accreted terranes, contains evidence for crustal thickening contemporaneous with horizontal shortening in the adjacent foreland. Basement contraction, resulting in crustal thickening and decreasing width of the continental margin west of the craton-shelf hinge, is localized in areas underlain by previously deformed transitional continental crust. Contraction of cover strata occurs at and east of the craton-shelf hinge. Shortening of deep crustal levels in the western portion of the continental margin and crustal thickening may balance cover shortening in the foreland to the east. Calculations, based on foreland displacement magnitude and crustal thickening to the west demonstrate the feasibility of such a kinematic connection between these areas.

Important conclusions of this study are that displacements within the continental margin are both horizontally and vertically distributed, that variations in horizontal displacements in the foreland correlate with variations in crustal thickening to the west, and that deformation of basement rocks is restricted to areas of previously deformed transitional continental crust: the craton maintains its integrity.

The Continental Margin

The western North American cordillera consists of two fundamentally different groups of rocks, those that are and always have been part of the North American continent and those that are partially or entirely exotic to the continent. The first group occupies an inboard position with respect to the second and includes continental crust, transitional crust, and sediment deposited thereupon and generally derived from the continent. The second group lies athwart the present plate boundary and includes oceanic and island arc basement and affiliated sediments as well as pieces of continental material of unknown derivation.

The exotic terranes accreted to North America at various times during the Phanerozoic and represent far-traveled masses with deformational and thermal histories distinct from the North American continent and from one another. The motion of these terranes can only be inferred from displaced faunal assemblages, paleomagnetic anomalies and long-distance lithic correlations. The complex and divergent histories of these terranes coupled with the difficulty of determining their origin, plate tectonic setting and displacement path precludes a simple model of the deformational processes involved.

In contrast, rocks that have always been part of the North American continent, although often allochthonous, are not far-traveled and display marked similarity in deformational and thermal history. The ability to infer the original position of these rocks and their historical affiliation with a single plate allow analysis of the kinematics of continental margin deformation. Therefore, we make a distinction between the morphologic edge of the North American plate and the continental margin defined herein as the western limit of continental crust of North America, sediments deposited on continental North America and rocks initially or secondarily modified or derived from the North American continent.

As so defined, the continental margin of western North America was initiated during a late Proterozoic rifting event in which a continental fragment of unknown size drifted away from the newly formed

North American continental margin. West of a hinge zone, the continental crust that had been thinned and faulted during the rifting event subsided and accumulated sediment. Thus, by the beginning of Phanerozoic time a passive continental margin occupied western North America. Although accretion of terranes to portions of the western margin of North America occurred during Paleozoic time, the passive margin persisted until at least the Early Triassic when transition to the active margin extant today occurred. Paleozoic accretion did not result in pervasive deformation of the continental margin and only locally interrupted passive margin sedimentation.

Following formation of an active margin, and predominantly in the Jurassic and Cretaceous, the majority of exotic terranes accreted to western North America and the continental margin underwent pervasive deformation that extended well east of the Paleozoic hinge and permanently terminated shelf sedimentation. Due to the accretion of terranes of various size at different times and positions the morphologic edge of North America grew sporadically westward accompanied by a westward shifting plate boundary. Although terrane accretion was distributed both temporally and spatially along the western margin of North America, and, thus, the history of accretion varies as a function of position, the deformation of the continental margin is remarkably similar over large areas and occurred in roughly synchronous pulses. The suggestion that accretion of terranes was not the cause of continental deformation and that both may have been manifestations of the same plate margin phenomena justifies an examination of continental margin deformation exclusive of accreted terranes.

The continental margin, as defined above, is bounded to the west by tectonic contacts with accreted terranes. To the east, the continental margin merges smoothly into the cratonic heart of North America. The continental margin can be divided into two distinct provinces (Fig. 1). A foreland, consisting of imbricated and folded sheets of continentally derived platform and shelf strata, occurs inboard of a hinterland composed of plutonic rocks, metamorphic rocks of varied age and protolith, and continentally affiliated strata of the outer shelf and adjacent oceanic basin. Common usage often includes accreted terranes within the hinterland; we refer to the belt of variable width between the foreland and the eastern limit of accreted terranes as the continental hinterland.

The Foreland Fold and Thrust Belt

Extending continuously from the southern United States to northern Canada, the Cordilleran foreland thrust belt is one of the most continuous geologic provinces in the World. The eastern or inboard boundary of the continental margin is defined at the western limit of undeformed platformal cover of the North American craton. Various structural geometries occur at the transition between the thrust belt and undeformed craton, but a synclinal trough is often developed at the eastern deformation front of the thrust belt and may be paired with an anticlinal arch farther east. The synclinal trough may result from flexure of the continental crust beneath the load of emplaced thrust sheets (Jordan, 1981; Lorenz, 1982), from monoclinal draping of cover strata above the termination of a blind thrust (Woodward, 1981, 1982), or from tectonic wedging of the thrust belt into the platformal cover and formation of a triangle zone (Price, 1981, 1986; Jones, 1982). The eastern margin of the foreland thrust belt describes several salients and recesses (Fig. 1) that are geographically related to the presence of foreland basement uplifts or paleogeographic declivities in the continental margin.

In the central United States segment of the Cordillera thrust belt, the eastern boundary of the deformed belt can not be determined by the juxtaposition of deformed and undeformed strata because Laramide Rocky Mountain foreland basement uplifts overlap or coincide with the eastern margin of thrust belt deformation (Beutner, 1977; Perry and others, 1983). The position of the deformation front and relationship between the thrust belt and uplifts have been clarified by seismic reflection profiles and drill holes (Perry and others, 1983; Lopez and Schmidt, 1985; Blackstone, 1977; Dixon, 1982) suggesting that foreland basement uplifts have little influence on the geometry of displacement termination, but apparently influence the orientation of thrust belt structures (Brandon, 1984; Perry and others, 1983). They may also influence the position, timing and spacing of thrusts (Kopania, 1985; Blackstone, 1977) although these effects are not universally accepted (Woodward, 1986).

The salient in the thrust belt in west-central Montana, the only major irregularity in the eastern margin of the thrust belt north of the zone of foreland basement uplifts, is coincident with an embayment in the Proterozoic continental margin that is the Belt basin (Harrison and others, 1974). Bounded to the south by the Lemhi arch (Ruppel, 1985), Salmon River arch (Armstrong, 1975), and the Dillon basement block (Gieger, 1985), this Proterozoic embayment in the continental margin allowed the thrust belt deformation front to migrate well east of its position to the north and south (Woodward, 1981, 1982).

The thrust belt deformation front has a smooth arcuate trace where not affected by irregularities in the pre-existing continental margin or foreland basement uplifts. Therefore, as suggested by Beutner (1977) more than a decade ago, the sinuous trace of the eastern margin is not an inherent feature of the thrust belt but rather an inherited one.

The thrust belt, in addition to its length and continuity, is notable for its remarkably consistent structural geometry. Many of the generally accepted "rules" of thrust belt development have their origin in studies of the North American Cordillera (Boyer and Elliott, 1982; Royse and others, 1975; Price, 1981; Dahlstrom, 1977; Allmendinger, in press). At the surface, the foreland thrust belt is characterized by imbricate thrust sheets, fault accommodation folds, and cataclasis on fault surfaces. In addition, it has been observed that thrust faults cut upsection in the direction of displacement and, with a few notable exceptions (Allmendinger, 1981; Royce, 1985; Lamerson, 1982), develop from west to east such that the oldest thrust is the westernmost and structurally highest thrust in the belt whereas, the easternmost and structurally lowest thrust is youngest (Armstrong and Oriel, 1965; Price and Mountjoy, 1970; Royse and others, 1975; Price, 1981; Woodward, 1981, 1986; Allmendinger, 1981; Perry and Sando, 1982; Lamerson, 1982). Seismic reflection profiling and drilling within the foreland thrust belt have shown that in the subsurface major thrust faults have ramp and flat geometry and sole into a basal decollement (Dixon, 1982; Royse and others, 1975; Perry and others, 1983; Allmendinger and others, 1986). Although the basal decollement occurs within the platform cover sequence locally, and generally climbs to higher stratigraphic levels from west to east across the foreland, the decollement is at the contact between platformal cover strata and crystalline basement throughout the majority of the foreland in Canada (Price, 1981), Montana (Harrison and others, 1980), and Wyoming and Utah (Royse and others, 1975; Dixon, 1982). These characteristics lead to similar structural style throughout the foreland thrust belt and imply that the belt records about 50% shortening overall (Royse and others, 1975). Moreover, individual thrust sheets contain similar stratigraphic packages and undergo increasing total displacement from east to west. Although these characteristics do not all exist at each position within the foreland thrust belt, together they uniquely define the belt.

The Foreland-Hinterland Transition

The western margin of the foreland thrust belt, the transition from foreland to continental hinterland, is a diffuse and variably defined zone. The continental hinterland contains crystalline basement rocks in thrust sheets or nappes, plutons with chemistry indicative of continental crustal derivation or contamination, and structures of variable vergence and propagation history. The continental hinterland is characterized by more ductile deformation than the foreland, indicated by recumbent isoclinal folds, sheath folds and mylonitized fault zones. Although these characteristics together define the hinterland as a province distinct from the foreland and allow the surface trace of the transition to be approximately located, none individually defines the boundary between foreland and hinterland everywhere. For example, plutons such as the Boulder Batholith (Hyndman and others, 1975; Hyndman, 1979; Hamilton and Myers, 1974) occur in areas with character more typical of foreland than hinterland and crystalline basement is not always present near the transition.

At the surface, the transition from foreland to hinterland commonly occurs across an anticline-syncline pair of multikilometer amplitude. In Canada and northern Montana, the Purcell anticlinorium is flanked to the west by the Kootenay arc across a large-amplitude monoclinal step (Price, 1981; Monger and others, 1985). In western Utah and eastern Nevada, anticlinal nappe piles associated with the Willard-Paris thrust system (Bruhn and Beck, 1981; Schirmer, 1985) occur east of the Sublet and Confusion Range synclines (Armstrong, 1982). In both areas, slices of crystalline basement rock occur within the anticlinoria. The Malton gneiss and several smaller gneiss bodies are in the hangingwall to the Purcell thrust in the Selwyn Range of British Columbia (Mountjoy and Forest, 1986; Leonard, 1984; McDonough and Simony, 1984) and the Farmington complex occupies a similar position with respect to an anticlinal nappe pile in central Utah (Bruhn and Beck, 1981). Crystalline basement also occurs in thrust sheets near the Idaho-Montana border southeast of the Idaho batholith (Skipp, 1985). With the exception of west-central Montana, the transition from foreland to hinterland also marks the eastern limit of abundant plutonic rocks. The foreland-hinterland boundary at the surface is, therefore, marked by a large west-facing monocline, involvement of crystalline rocks in thrust sheets, and the occurrence of plutons.

Abundant plutonism and basement involvement in thrust sheets imply that the transition from foreland to hinterland represents a boundary between areas underlain by undeformed crystalline basement and areas underlain by deformed basement. In Canada, the western margin of undeformed basement has been located by the truncation of a magnetic fabric associated with basement rocks beneath the foreland and craton (Price, 1981; Mongerand others, 1985). The truncation occurs near the large-amplitude monoclinal step that marks the surface trace of the hinterland-foreland boundary. In southern British Columbia, northern Washington and Idaho, and Utah and Nevada, deep seismic reflection profiles (Allmendinger and others, 1986; Hauser and others, 1987; Potter and others, 1986; Klemperer and others, 1987) have revealed that the hinterland is underlain by basement rocks that exhibit a seismic fabric or are entirely non-reflective, whereas, the foreland and craton are underlain by basement characterized by diffuse diffractions. The difference in seismic fabric may indicate a transition from undeformed to deformed crystalline basement. Provided that the change in seismic character of deep crustal levels and the truncation of magnetic fabrics represent the western limit of undeformed basement the hinterland-foreland boundary can be located geophysically at depth.

In two widely spaced COCORP seismic reflection profiles (Potter and others, 1986; Allmendinger and others, 1986), west-dipping reflectors continuous to deep crustal levels are present and offset by deeply penetrating east-dipping reflectors within the continental hinterland. Although the west-dipping reflectors may owe their orientation to rotation on east-dipping surfaces and may be entirely Cenozoic features, the possibility remains that these reflectors are zones of displacement transfer from deep crustal levels in the hinterland to shallow levels in the foreland (Allmendinger and others, 1986). It is notable that the west-dipping reflectors in Nevada and Utah occur beneath the surface trace of the hinterland-foreland boundary (Allmendinger and others, 1986).

The Continental Hinterland

The width of the continental hinterland at depth between the zone of displacement transfer to the foreland and the western limit of continental basement varies considerably along the strike of the Cordillera (Fig. 1). The continental hinterland is approximately 550km wide across western Utah and eastern Nevada, but narrows to less than 100km in west-central Idaho. The continental hinterland widens to

>300km in northern Washington and narrows again to the north where the Shuswap metamorphic complex and the continental hinterland are roughly coextensive. These variations in the width of the continental hinterland correlate at different positions with different amounts of Tertiary extension, salients in the Cretaceous magmatic arc and declivities in the inherited continental margin. We suggest below that the variation in continental hinterland width also correlates with the amount of shortening in the adjacent segment of the foreland thrust belt.

Preserved cover strata in the continental hinterland rarely include the upper portion of the stratigraphic section. Where such strata are present they are little deformed. In Eastern Nevada regional concordance of upper Paleozoic and Tertiary strata (Armstrong, 1972; Gans and Miller, 1983) demonstrates the lack of deformation at high stratigraphic levels. The vergence of structures, overwhelmingly eastward in the foreland thrust belt, varies across the continental hinterland. Variable vergence is represented by large fan structures in Canada (Brown and Tippett, 1978; Pell and Simony, 1984; Garwin and others, 1987; Dechense and others, 1984) and reported from the Snake Range and eastern continental hinterland of Utah (Miller and others, in press; Allmendinger and Jordan, 1984; Allmendinger and others, 1985). West-verging structures including, thrust faults (Elison, 1987), kilometer scale folds (Hyndman, 1979), and shear zones (Murphy, 1985; Brown and others 1986), occur preferentially near the western margin of the continental hinterland.

The continental hinterland contains multiple plutons and most of the belt of metamorphic core complexes from the Shuswap metamorphic complex in British Columbia to the Snake Range of Nevada (Armstrong, 1982). In the United States, the Cretaceous magmatic arc crosses the western edge of the continental hinterland from an outboard position in the southern and central United States and in Canada to a position within the continental hinterland in western Montana and Idaho. Where Tertiary extension has exposed high grade metamorphic rocks, the contact between high grade rocks and overlying strata is often a mylonitic detachment (Rhodes and Hyndman, 1984; Snoke and Miller, 1988) and in some cases this detachment is demonstrably Mesozoic (Bickford and others, 1985; Journeay and Brown, 1986).

Analogous to the foreland where details vary, but the thrust belt as a whole displays a characteristic structural geometry, the continental hinterland is similar in general character for at least 1500km along its north-south trend. Similarities of different portions of the continental hinterland include; variably vergent structures in cover strata with large-scale west-verging structures, little deformation of shallow stratigraphic levels where exposed, abundant plutonism and metamorphism, and occurrence of mylonitic detachment zones.

The continental hinterland is bounded to the west by tectonic contacts with accreted terranes. At the surface this contact is marked by the juxtaposition of continentally derived shelf strata or strata transitional between the passive margin shelf on the east with basinal strata, volcanic arc rocks, arc platform and fringe sediments, and oceanic crust and cover on the west. The time of attachment of the easternmost terrane to North America varies from Mississippian in Nevada and central Idaho (Silberling and Roberts, 1962; Isaacson and McFaddan, 1984; Dover, 1980; Davis, 1984; Marshall and Myers, 1984; McFaddan, 1985), and possibly also in parts of British Columbia (Gehrels and Smith, 1987; Gordey and others, 1987; Mortensen and others, 1987), to Cretaceous in western Idaho (Hyndman, 1979; Hyndman and Alt, 1987; Sutter and others, 1984). At depth, the western limit of the continental hinterland is defined by the isotopic systematics of plutonic rocks presumed to sample deep crustal levels. Variations in initial isotopic ratios allow determination of the westernmost plutons that were derived from or contaminated by continental crustal material. According to arguments by Kistler and Peterman (1973, 1978) rocks with $^{87}Sr/^{86}Sr > 0.706$ initial ratios indicate derivation from or contamination by continental material. The $^{87}Sr/^{86}Sr = 0.706$ initial ratio isopleth (Fig. 1) has been delimited in Nevada (Kistler, 1983; in press), in Oregon, Washington, and Idaho (Armstrong and others, 1977; Babcock and others, 1985), and in southern British Columbia (Parrish, 1981; Ewing, 1981). The edge of continental crust as defined by Sr isotopes is generally collaborated by other geochemical characteristics (Farmer and Depaolo, 1983; Lee and others, 1981; Miller and Bradfish, 1980).

West of the continental margin, crystalline basement is non-continental in isotopic character and probably affiliated with accretionary events (see for example Speed, 1979; Monger, 1985; Monger and others, 1985). The western limit of the continental margin as defined at the surface and at depth generally coincide. Accreted terranes extend well east of the isotopically defined western edge of continental basement in only two areas. In Nevada, the Roberts Mountains allochthon extends roughly 100km inboard from the $^{87}Sr/^{86}Sr = 0.706$ line (Speed et al, in press). Exposures of near shore carbonate strata beneath the Roberts Mountains allochthon at the $^{87}Sr/^{86}Sr = 0.706$ line in north-central Nevada (Elison, 1987) document the original western extent of Paleozoic shelf strata. In northern Washington and southern British Columbia, where accreted terranes also extend inboard of the western continental margin, a similar situation has been inferred (Halwas and Simony, 1986; Orr, 1985). Accreted terranes extending east of the edge of continental basement were detached from their basement and thrust across the outer shelf strata and subjacent transitional continental crust. The position of the boundary between continent and accreted terranes is, thus, diffuse and varies with depth.

Displacement Timing, Magnitude and Rate in the Foreland Thrust Belt

The existence of paleogeographic or tectonic irregularities that correlate with each of the salients or recesses in the eastern margin of the thrust belt suggests that the salients and recesses are not related to variations in the amount of displacement along the trend of the thrust belt. This conclusion is supported by estimates of displacement for several segments of the Cordilleran thrust belt (Table 1). These estimates suggest a continuous decrease in total foreland thrust belt displacement from north to south independent of the position of salients in the deformation front. Total displacement varies smoothly along the thrust belt, but shortening strain is heterogeneous due to pre-existing irregularities (for example, Woodward, 1986). The implication is that the displacement magnitude is independent of the pre-existing geometry of the continental margin, but that the location and magnitude of shortening are strongly influenced by that geometry.

The time span during which foreland thrust belt deformation occurred probably varies along the thrust belt. Dating the time at which deformation ceased is straightforward and relatively precise. On the other hand, constraining the age of onset of foreland thrusting has been more difficult and in some cases controversial (compare Wiltschko and Dorr, 1983; Heller and others, 1986).

Overlapping strata constrain the frontal portion of the thrust belt in Canada and northern Montana to have ceased displacement by late Eocene or early Oligocene time (Price, 1981; Bally, 1984). Similarly, igneous dikes that cross-cut the Steinbach and Eldorado thrusts in the eastern part of the northern

Montana thrust belt give K/Ar ages of 46.3 and 58.3 Ma. (Mudge and Earhart, 1980). The youngest deformed strata in northern Montana and Canada constrain the McConnel and Lewis thrusts to be younger than latest Cretaceous (Price, 1981; Winston and Woods, 1986).

Perhaps the best technique for determining the history of the foreland thrust belt at any position is analysis of foreland basin subsidence and sedimentation. The subsidence history of foreland basins is theoretically related to the history of loading by thrust sheets (Jordan, 1981; Speed and Sleep, 1982) and thus the time of basin subsidence should relate to the time of thrusting. Moreover, uplift inherent in thrust sheet emplacement provides a source of coarse material to the foreland basin and dating the stratigraphic succession of synorogenic conglomerates provides estimates of the times of thrusting. Basin subsidence curves have been used to constrain the age of thrusting in the Idaho-Wyoming-Utah thrust belt (Heller and others, 1986) and foreland basin stratigraphy has provided further constrain there (Royse and others, 1975; Wiltshcko and Dorr, 1983; Heller and others, 1986) and in British Columbia (Thompson, 1979; Price, 1981).

A method for dating the motion on individual thrusts is to radiometrically date, by the K/Ar method, illites and potash bentonites that form in strata overridden by thick thrust sheets (Hoffman and others, 1976; Aronson and Elliott, 1985). Because the minerals dated were formed by imposition of the thrust sheet, they provide constraints on the timing of thrust motion.

Given the above constraints, it is possible to estimate the time period of thrust belt formation in different segments of the thrust belt and derive the apparent rate of displacement for these segments. Deformation periods for the entire thrust belt and smaller portions or individual thrusts where available are shown in Table 1 along with the derived displacement rates. It is apparent from this table that the displacement in the foreland thrust belt of western North America ceased by about 50 to 55mybp. The table suggests that the onset of displacement varied by over 50 million years between about 150mybp and 100mybp. A dominant feature of the displacement rate calculations is that rates calculated for shorter time periods are invariably higher than rates for the thrust belt as a whole at the same location. This pattern may represent sporadic periods of thrust development and displacement interspersed with periods of quiescence throughout a segment of the thrust belt. Sporadic thrusting is supported by evidence from foreland basins in British Columbia, Utah, and Wyoming (Thompson, 1979; Heller and others, 1986; Wiltshkco and Dorr, 1983) where pulses of synorogenic conglomerate deposition are separated by times of little sedimentation. If these pulses of sedimentation correspond to pulses of thrusting and if times of little or no coarse clastic sedimentation record times of no thrusting, then the time period over which active thrust displacement occurred is reduced and the rate of displacement for the entire foreland thrust belt approaches the rate of its parts. The difference between individual thrust displacement rates and thrust belt displacement rates provides an estimate of the relative length of active and quiescent periods.

Variations in active displacement rates through time are evident in the Idaho-Wyoming-Utah segment of the thrust belt where individual thrusts apparently displaced at different rates; the youngest thrusts generally displacing the most rapidly (Table 1). Sporadic thrust displacement is also indicated by the low displacement rate of the entire belt relative to the rates of the individual thrusts. Some of the variability in displacement rates is undoubtably a result of poorly constrained displacement period. This is particularly evident in the case of the Willard/Paris thrust that has 0.55 mm/yr to >1.5 mm/yr of displacement dependent on the interpretation of displacement onset age.

Displacement Timing, Magnitude and Rate in the Continental Hinterland

The magnitude of horizontal displacement within the continental hinterland is difficult to constrain in part because of the complexity of structures of variable vergence and in part because of the masking effects of metamorphism and plutonism. Transcurrent and extensional deformation that occurred synchronously or later than compressional deformation at various positions within the Cordillera, and Paleozoic accretionary events that deformed the outer portions of the hinterland prior to the onset of major contraction have created structures that complicate analysis of contractional displacement. The hinterland, however, contains abundant plutonic and metamorphic rock and radiometric dating of deformed and undeformed crystalline rocks allows the period of displacement to be constrained. Furthermore, the magnitude of vertical displacements can be better constrained within the hinterland than within the foreland because mineralogic assemblages in metamorphic terranes allow the temperature-pressure conditions of metamorphism to be determined. Recent studies of metamorphic core complexes (Snoke and Miller, 1988; Miller and others, 1988; Journeay and Brown, 1986; Dallmeyer and others, 1986) have demonstrated Mesozoic tectonic, metamorphic, and plutonic histories in many of these complexes. Table 2 gives depth and age relations for metamorphic rocks, plutons and mylonite zones affiliated with the metamorphic core complexes and surrounding areas of the continental hinterland.

The time period over which metamorphism, deformation, and plutonism occurred in the continental hinterland varies along the trend of the North American Cordillera. Jurassic metamorphism is known from several core complexes (Dallmeyer and others, 1986; Archibald and others, 1983) and Jurassic plutonism and deformation are also prevalent in the hinterland of Canada and Utah and Nevada (Allmendinger and others, 1985; Journeay and Brown, 1986; Parrish and Wheeler, 1983; Miller and others, 1988; Parkinson, 1985; Fox and others, 1977). No evidence for Jurassic plutonism, metamorphism or deformation has been found in the central Idaho portion of the hinterland. Cretaceous plutonism and deformation are distributed along the entire length of the continental hinterland (Armstrong and Suppe, 1973; Moore and McKee, 1983; Miller and Engels, 1975; Parkinson, 1985; Silberman and McKee, 1971) and Tertiary metamorphism and uplift associated with metamorphic core complexes (Howard, 1971; 1980; Snoke, 1980; Miller, 1980; Dallmeyer and others, 1986; Dokka and others, 1986; Parkinson, 1985; Corbett and Simony, 1984; Halwas and Simony, 1986; Read and Brown, 1981) has long been recognized.

The rates of vertical displacement of hinterland rocks vary considerably, in part due to poorly constrained displacement period. Vertical displacement during the late Jurassic through late Cretaceous during primarily contractile deformation apparently occurred at higher rates than during Tertiary extension. Moreover, vertical displacement rates in the continental hinterland are less than rates of horizontal displacements in adjacent foreland regions. Modern models of continental margin deformation generally consider that the displacement of cover strata above undeformed continental crust in the foreland is balanced by deformation of transitional crust in the hinterland (Brown and others, 1986; Monger and others, 1985; Speed and others, 1988; Allmendinger, in press). Although these models vary somewhat in detail, a salient point is that horizontal shortening and concomitant thickening occurs in hinterland basement that is detached from cover strata. Cover strata shortens farther to the east in the foreland fold and thrust belt. Given the current width of the continental hinterland between the eastern limit of accreted terranes and the zone of displacement transfer to the foreland and given the magnitude of horizontal displacements within the foreland, the amount of hinterland thickening that would result from balanced shortening can be estimated. If transitional crust deformed in plane strain, the amount of crustal thickening (T_h) necessary to balance a given amount of foreland shortening (S_f) is given by

$$1.0 + S_f/W_h = T_h,$$

where W_h is the current width of the continental hinterland and T_h is the ratio of the deformed hinterland crustal thickness to the initial hinterland crustal thickness. Table 3 gives predicted values of hinterland thickening for various positions along the western North American Cordillera. Not surprisingly, the predicted thickening decreases from north to south along the Cordillera, in part due to decreasing amounts of foreland shortening and in part due to increasing width of the continental hinterland. Obviously, Tertiary extension, most prominent in the Great Basin, but known to extend along the entire continental hinterland from southern Nevada to British Columbia, has altered the post-contractional width of the continental hinterland. This extension was taken into account in the calculation of predicted shortening (Table 3). Amounts of Tertiary extension at various points along the western Noth American Cordillera were taken as follows: 20% across southern British Columbia (Okulitch, 1985), 40% across northern Washington and Idaho, 70% across the northernmost Great Basin at 42° N (Wells and Heller, 1988), and 100% across the central Great Basin at 40° N (Speed and others, 1988; Bogen and Schweickert, 1985). The north to south increase in Tertiary extension is demonstrated by rotations of outboard terranes determined paleomagnetically (Wells and Heller, 1988; Bogen and Schweickert, 1985). Removing the values of extension cited above from the current width of the continental hinterland results in more uniform continental hinterland width along the cordillera.

DISCUSSION

It is evident from Table 3 that the amount of crustal thickening predicted for British Columbia and for the Great Basin are not unreasonable. In both locations, crystalline basement is locally exposed at the surface and seismic reflection profiles indicate crustal thicknesses of 30 to 40 km. The implication is that it is conceivable that foreland shortening of cover strata is accommodated by shortening of basement in the hinterland.

Table 3 shows only the predicted magnitude of crustal thickening in the continental hinterland. The rate at which this thickening proceeds, relative to horizontal shortening, however, is dependent on the kinematics of deformation. If the transitional continental crust in the hinterland shortens and thickens by thrust imbrication, then the relative magnitude of shortening and thickening are related by the dip of the imbricating thrusts by

$$T_h = S_h \tan\alpha,$$

where T_h is defined above, S_h is the magnitude of hinterland shortening, and α is the dip of the thrusts. Because motion on the thrusts produces thickening and shortening concurrently, the rates of thickening and shortening are related by the same proportion. For thrust dip angles between 5 and 25 degrees, the ratio of thickening to shortening and of thickening rate to shortening rate vary from 0.087 to 0.467. Deformation of the continental hinterland by thrust imbrication has been suggested for the southern British Columbia hinterland (Monger and others, 1985; Journeay and Brown, 1986; Brown and others, 1986). Thickening by thrust imbrication provides a mechanism to expose high grade metamorphic mineral assemblages at high stratigraphic levels.

If, on the other hand, transitional continental crust deforms in homogeneously distributed pure shear with horizontal contraction and vertical extension, the relative magnitudes and rates of thickening are dependent on the initial thickness of transitional crust and the width across which thickening obtains by

$$(1.0\text{-}T_h)/S_h = t_i/W_h,$$

where t_i is the initial thickness of hinterland basement and W_h, is the deformed width of the continental hinterland. For hinterland aspect ratios > 5.0, the ratio of thickening to shortening and thickening rate to shortening rate are < 0.2. Speed and others (1988) have suggested a model along these lines for the continental hinterland of the northern Great Basin. In eastern Nevada, where high grade metamorphism has effected rocks as young as Permian (Snoke, 1980; Snoke and Miller, 1988), some thrust imbrication is necessary. Farther west, rocks as old as Proterozoic are little metamorphosed and basement thickening without imbrication is admissible.

The amount of crustal thickening predicted for western Montana and east-central Idaho, given the restricted width of continental hinterland, appears unrealistically high. This area also differs from the continental margin to the north and south in that it is characterized by both plutonism and Tertiary extension on west-dipping normal faults extending well east of the foreland-hinterland boundary defined at the surface. Moreover, a Proterozoic embayment in the continental margin, the Belt basin, occupied this region. The deformation of crystalline basement in this portion of the North American cordillera probably extends farther east than elsewhere, based on the plutonism and extensional deformation, and may represent the effects of the embayment in the continental margin that contains stretched or faulted transitional crust below the Belt basin.

Conclusions

It is concluded that shortening in cover strata recorded by the foreland thrust belt may be accommodated by shortening and thickening of crystalline basement in the continental hinterland. This conclusion requires that displacement within the continental margin is inhomogeneously distributed both horizontally and vertically and requires detachment of the basement and cover. Furthermore, detachment of hinterland basement from crystalline basement beneath the foreland is also required. The coincidence of the foreland-hinterland boundary and the craton-shelf hinge, the concentration of plutonism and metamorphism in the hinterland, the apparent restriction of basement deformation to the hinterland and the restriction of post-contractile extension to the area of previously deformed basement all argue that the existence and behavior of transitional continental crust influence the distribution of continental margin deformation. Deformation in basement only occurs in areas underlain by continental crust deformed in extension and/or contraction. The craton does not deform at depth and provides a buttress against which strata in the foreland thrust belt deforms.

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REFERENCES CITED

Allmendinger, R. W, 1981, Structural geometry of Meade thrust plate in northern Blackfoot Mountains, southeastern Idaho.: American Association of Petroleum Geologists Bulletin, v. 65, p. 509-525.

Allmendinger, R. W., Farmer, H., Hauser, E., Sharp, J., Von Tish, D., Oliver, J., and Kaufman, S., 1986, Phanerozoic tectonics of the Basin and Range-Colorado Plateau transition from COCORP data and geologic data: a review., Reflection Seismology: The Continental Crust Geodynamics Series Vol. 14., American Geophysical Union, p. 257-267.

Allmendinger, R. W., and Jordan, T. E., 1981, Mesozoic evolution, hinterland of the Sevier orogenic belt.: Geology, v. 9, p. 308-314.

—— 1984, Mesozoic structure of the Newfoundland Mountains, Utah: horizontal shortening and subsequent extension of the hinterland of the Sevier Belt.: Geological Society of America Bulletin, v. 95, p. 1280-1292.

Allmendinger, R. W., Miller, D. M., and Jordan, T. E., 1985, Known and inferred Mesozoic deformation in the hinterland of the Sevier Belt, Northwest Utah., *in* Geology of Northwest Utah, Southern Idaho, and Northeast Nevada, Kerns, G. J., ed., Utah Geological Association, Pub. no. 13, p. 21-34.

Archibald, D. A., Glover, J. K., Price, R. A., Farrar, E., and Carmichael, D. M., 1983, Geochronology and tectonic implications of magmatism and metamorphism, southern Kootenay arc and neighboring regions, southeastern British Columbia. Part 1: Jurassic to mid-Cretaceous.: Canadian Journal of Earth Science, v. 20, p. 1891-1913.

Armstrong, F. C., and Oriel, S. S., 1965, Tectonic development of Idaho-Wyoming thrust belt.: American Association of Petroleum Geologists Bulletin, v. 49, p. 1847-1866.

—— 1972, Low-angle (denudation) faults, hinterland of the Sevier orogenic belt, eastern Nevada and western Utah.: Geological Society of America Bulletin, v. 83, p. 1729-1754.

—— 1975, Precambrian (1500my old) rocks of central Idaho - the Salmon River Arch and its role in Cordilleran sedimentation and tectonics.: American Journal of Science, v. 275-A, p. 437-467.

—— 1982, Cordilleran metamorphic core complexes - from Arizona to southern Canada.: Annual Reviews of Earth and Planetary Sciences, v. 10, p. 129-154.

Armstrong, R. L., and Suppe, J., 1973, Potassium-Argon geochronometry of Mesozoic igneous rocks in Nevada, Utah and southern California.: Geological Society of America Bulletin, v. 84, p. 1375-1392.

Armstrong, R. L., Taubeneck, W. H., and Hales, P. O., 1977, Rb-Sr and K-Ar geochronometry of Mesozoic granitic rocks and their Sr isotopic composition, Oregon, Washington, and Idaho.: Geological Society of America Bulletin, v. 88, p. 321-331.

Aronson, J., and Elliott, W., 1985, Dating of thrusting in the Idaho-Wyoming belt: K-Ar ages of authigenic illite formed by burial beneath thrust sheets and or synorogenic sediments.: Geological Society of America Abstracts with Programs, v. 17, p. 207.

Babcock, R. S., Armstrong, R. L., and Misch, P., 1985, Isotopic constraints on the age and origin of the Skagit metamorphic suite and related rocks.: Geological Society of America Abstracts with Programs, v. 17, p. 339.

Bally, A. W., 1984, Tectogenese et sismique reflexion: Bulletin Societe Geologique de France, v. 26, p. 279-285.

Berger, B. R., and Taylor, B. E., 1980, Pre-Cenozoic normal faulting in the Osgood Mountains, Humboldt County, Nevada.: Geology, v. 8, p. 594-598.

Beutner, E. C., 1977, Causes and consequences of curvature in the Sevier orogenic belt, Utah to Montana., *in* Rocky Mountain thrust belt geology and resources, Heisey, E. L., Lawson, D. E., Norwood, E. R., Wach, P. H., and Hale, L. A., eds., Wyoming Geological Association 29th Annual Field Conference, p. 353-365.

Blackstone, D. L., Jr., 1977, The overthrust belt salient of the cordilleran fold belt, western Wyoming - southeastern Idaho - northeastern Utah., *in* Rocky Mountain thrust belt geology and resources, Heisey, E. L., Lawson, D. E., Norwood, E. R., Wach, P. H., and Hale, L. A., eds., Wyoming Geological Association 29th Annual Field Conference, p. 367-384.

Bogen, N. L., and Schweickert, R. A., 1985, Magnitude of crustal extension across the northern Basin and Range province: constraints from paleomagnetism.: Earth and Planetary Science Letters, v. 75, p. 93-100.

Boyer, S. E., and Elliott, D., 1982, Thrust systems.: American Association of Petroleum Geologists Bulletin, v. 66, p. 1196-1230.

Brandon, W. C., 1984, An origin for the McCartney's Mountain salient of the southwestern Montana fold and thrust belt.: Geological Society of America Abstracts with Programs, v. 16, p. 216.

Brown, R. L., Journeay, M., Lane, L. S., Murphy, D. C., and Rees, C. J., 1986, Obduction, backfolding, and piggyback thrusting in the metamorphic hinterland of the southeastern Canadian Cordillera.: Journal of Structural Geology, v. 8, p. 255-268.

Brown, R. L., and Tippett, C. R., 1978, The Selkirk fan structure of the southeastern Canadian Cordillera.: Geological Society of America Bulletin, v. 89, p. 548-558.

Bruhn, R. L., and Beck, S. L., 1981, Mechanics of thrust faulting in crystalline basement, Sevier orogenic belt, Utah.: Geology, v. 9, p. 200-204.

Chase, R. B., Bickford, M. E., and Arruda, E. C., 1983, Tectonic implications of Tertiary intrusion and shearing within the Bitterroot Dome, northeastern Idaho Batholith.: Journal of Geology, v. 91, p. 462-470.

Chesson, S. A., Billodeau, W. L., and Wells, J. D., 1984, Cambrian stratigraphy of the Libby and Alberton areas, northwestern Montana., *in* Northwest Montana and Adjacent Canada, McBane, J. D., and Garrison, P. B., eds., Montana Geological Society 1984 Field Conference and Symposium, p. 121-124.

Corbett, C. R., and Simony, P. S., 1984, The Champion Lake Fault in the Trail-Castlegar area of southeastern British Columbia.: Current Research Part A, Geological Survey of Canada Paper 84-1A, p. 103-104.

Dahlstrom, C. D. A., 1970, Structural geology in the eastern margin of the Canadian Rocky Mountains.: Bulletin of Canadian Petroleum Geology, v. 18, p. 332-406.

Dallmeyer, R. D., Snoke, A. W., and McKee, E. H., 1986, The Mesozoic-Cenozoic tectonothermal evolution of the Ruby Mountains, East Humboldt Range, Nevada: A cordilleran metamorphic core complex.: Tectonics, v. 5, p. 931-954.

Davis, S., 1984, Antler orogenic microstructures in the Devonian Milligan Formation, Pioneer Mountains, central Idaho.: Geological Society of America Abstracts with Programs, v. 16, p. 220.

Dechesne, R. G., Simony, P. S., and Ghent, E. D., 1984, Structural evolution and metamorphism of the southern Cariboo Mountains, near Blue River, British Columbia.: Current Research Part A, Geological Survey of Canada Paper 84-1A, p. 91-94.

Dixon, J. S., 1982, Regional structural synthesis, Wyoming salient of western overthrust belt.: American Association of Petroleum Geologists Bulletin, v. 66, p. 1560-1580.

Dokka, R. K., Mahaffie, M. J., and Snoke, A. W., 1986, Thermochronologic evidence of major tectonic denudation associated with detachment faulting, northern Ruby Mountains - East Humboldt Range, Nevada.: Tectonics, v. 5, p. 995-1006.

Dover, J. H., 1980, Status of the Antler Orogeny in central Idaho - clarifications and constraints from the Pioneer Mountains., *in* Paleozoic paleogeography of the west-central United States, Fouch, T. D., and Magathan, E. R., eds., Rocky Mountain Section Society of Economic Paleontologists and Mineralogists West-Central United States Paleogeography Symposium 1, p. 371-386.

Elison, M. W., 1987, Structural and tectonic history of the East Range, north-central Nevada; Implications for Mesozoic foreland tectonics. [Ph. D. Dissert.]: Evanston, IL., Northwestern Univ., 315 p.

Ewing, T. E., 1981, Regional stratigraphy and structural setting of the Kamloops Group, south-central British Columbia.: Canadian Journal of Earth Science, v. 18, p. 1464-1477.

Farmer, G. L., and DePaolo, D. J., 1983, Origin of Mesozoic and Tertiary granite in the western United States and implications for pre-Mesozoic crustal structure: I. Nd and Sr isotopic studies in the geosyncline of the northern Great Basin.: Journal of Geophysical Research, v. 88, p. 3379-3401.

Fox, K. F., Jr., Rinehart, C. D., and Engels, J. C., 1977, Plutonism and orogeny in north-central Washington - timing and regional context.: United States Geological Survey Professional Paper 989, 27 p.

Gans, P. B., and Miller, E. L., 1983, Style of mid-Tertiary extension in east-central Nevada.: Utah Geol. Min. Survey Special Studies 59, p. 107-139.

Garwin, S. L., Lewis, P. D., and Ross, J. V., 1987, Deformation and metamorphism in the N. Shuswap metamorphic complex, W. Cariboo Mtns., British Columbia.: Geological Society of America Abstracts with Programs, v. 19, p. 381.

Gehrels, G. E., and Smith, M. T., 1987, "Antler" allochthon in the Kootenay arc?: Geology, v. 15, p. 769-

770.

Geiger, B. C., 1985, Development of cleavage in the frontal fold and thrust belt, southwestern Montana: evidence for a foreland buttress.: Geological Society of America Abstracts with Programs, v. 17, p. 220.

Gordey, S. P., Abbott, J. G., Templeman-Kluit, D. J., and Gabrielse, H., 1987, "Antler" clastics in the Canadian Cordillera.: Geology, v. 15, p. 103-107.

Halwas, D. B., and Simony, P. S., 1986, The Castlegar Gneiss Complex, southern British Columbia: Current Research Part A, Geological Survey of Canada Paper 86-1A, p. 583-587.

Hamilton, W., and Myers, W. B., 1974, Nature of the Boulder Batholith of Montana.: Geological Society of America Bulletin, v. 85, p. 365-378.

Harrison, J. E., Griggs, A. B., and Wells, J. D., 1974, Tectonic features of the Precambrian Belt basin and their influence on poet-Belt structures.: United States Geological Survey Professional Paper 866, 15 p.

Harrison, J. E., Kleinkopf, M. D., and Wells, J. D., 1980, Phanerozoic thrusting in Proterozoic Belt rocks, northwestern United States.: Geology, v. 8, p. 407-411.

Hauser, E., Potter, C., Hauge, T., Burgess, S., Burtch, S, Mutschler, J., Allmendinger, R., Brown, L., Kaufman, S., and Oliver, J., 1987, Crustal structure of eastern Nevada from COCORP deep seismic reflection data.: Geological Society of America Bulletin, v. 99, p. 833-844.

Heller, P. L., Bowdler, S. S., Chambers, H. P., Coogan, J. C., Hogen, E. S., Shuster, M. W., Winslow, N. S., and Lawton, T. F., 1986, Time of initial thrusting in the Sevier orogenic belt, Idaho-Wyoming and Utah.: Geology, v. 14, p. 388-391.

Hoffman, J., Hower, J., and Aronson, J. L., 1976, Radiometric dating of time of thrusting in the disturbed belt of Montana.: Geology, v. 4, p. 16-20.

Howard, K. A., 1971, Paleozoic metasediments in the northern Ruby Mountains, Nevada.: Geological Society of America Bulletin, v. 82, p. 259-264.

—— 1980, Metamorphic infrastructure in the northern Ruby Mountains, Nevada., *in* Cordilleran Metamorphic Core Complexes, Crittenden, M. D., Coney, P. J., and Davis, G. H., eds., Geol. Soc.

America Memoir 153, p. 335-347.

Hyndman, D. W., 1979, Major tectonic elements and tectonic problems along the line of section from northeastern Oregon to west-central Montana.: Geological Society of America Map and Chart Series, Map MC-28C, 11 p.

Hyndman, D. W., and Alt, D., 1987, Late Cretaceous granite intrusions and thrust faulting, northern Rocky Mountains.: Geological Society of America Abstracts with Programs, v. x, x.

Hyndman, D. W., Talbot, J. L., and Chase, R. B., 1975, Boulder Batholith: a result of emplacement of a block detached from the Idaho Batholith infrastructure?: Geology, v. 3, p. 401-404.

Isaacson, P. E., and McFaddan, M. D., 1984, Silurian and Devonian history of central Idaho.: Geological Society of America Abstracts with Programs, v. 16, p. 225.

Jones, P. B., 1982, Oil and gas beneath east-dipping underthrust faults in the Alberta Foothills, Canada., *in* Geologic studies of the Cordilleran thrust belt, Powers, R. B., ed., Rocky Mountain Association of Geologists, p. 61-74.

Jordan, T. E., 1981, Thrust loads and foreland basin evolution, Cretaceous, western United States.: American Association of Petroleum Geologists Bulletin, v. 65, p. 2506-2520.

Journeay, M., and Brown, R. L., 1986, Major tectonic boundaries of the Omineca belt in southern British Columbia: a progress report.: Current Research Part A, Geological Survey of Canada Paper 86-1A, p. 81-88.

Kistler, R. W., 1983, Isotope geochemistry of plutons in the northern Great Basin., The role of heat in the development of energy and mineral resources in the northern Basin and Range province., Geothermal Resources Council Special Report 13, p. 3-8.

—— in press, Isotopes, igneous rocks, metamorphism, and tectonics in California and Nevada., *in* Rubey Volume 7, Metamorphism and Crustal Evolution, Western Coterminous United States, Ernst, G., ed.

Kistler, R. W., and Peterman, Z. E., 1973, Variations in Sr, Rb, K, Na, and initial Sr87/Sr86 in Mesozoic granitic rocks and intruded wall rocks in central California.: Geological Society of America Bulletin, v. 84, p. 3489-3512.

—— 1978, Reconstruction of crustal blocks of California on the basis of initial strontium isotopic composition of Mesozoic granitic rocks.: United States Geologic Survey Professional Paper 1071, 17 p.

Klemperer, S. L., Hauge, T. A., Hauser, E. C., Oliver, J. E., and Potter, C. J., 1986, The Moho in the northern Basin and Range province, Nevada, along the COCORP 40 N seismic reflection transect.: Geological Society of America Bulletin, v. 97, p. 603-618.

Kopania, A. A., 1985, The effects of basement-cored buttresses and major stratigraphic changes on thrust sheets: an example from the Idaho-Wyoming-Utah thrust belt.: Geological Society of America Abstracts with Programs, v. 17, p. 249.

Lamerson, P. R., 1982, The fossil basin area and its relationship to the Absaroka thrust system., *in* Geologic studies of the Cordilleran thrust belt, Powers, R. B., ed., Rocky Mountains Association of Geologists, p. 279-340.

Lee, D. E., Kistler, R. W., Friedman, I., and Van Loenen, R. E., 1981, Two mica granites of northeastern Nevada.: Journal of Geophysical Research, v. 86, p. 10607-10616.

Leonard, R., 1984, Metamorphism, structure, and stratigraphy around the Mount Blackman gneiss, British Columbia.: Current Research Part A, Geological Survey of Canada Paper 84-1A, p. 121-127.

Lopez, D. A., and Schmidt, C., 1985, Seismic profile across leading edge of fold and thrust belt of southwestern Montana.: American Association of Petroleum Geologists Bulletin, v. 69, p. 855.

Lorenz, J. C., 1982, Lithospheric flexure and history of the Sweetgrass Arch, northwestern Montana., *in* Geologic studies of the Cordilleran thrust belt, Powers, R. B., ed., Rocky Mountains Association of Geologists, p. 77-89.

Marshall, R. C., and Myers, P. B., Jr., 1984, Structural analysis of the Copper Basin Formation, central Pioneer Mountains, Idaho.: Geological Society of America Abstracts with Programs, v. 16, p. 246.

McDonough, M. R., and Simony, P. S., 1984, Basement gneisses and Hadrynian metasediments near Bulldog Creek, Selwyn Range, British Columbia.: Current Research Part A, Geological Survey of Canada Paper 84-1A, p. 99-102.

McFaddan, M. D., 1985, Silurian platform-margin transition in central Idaho: the Roberts Mountains Formation.: Geological Society of America Abstracts with Programs, v. 17, p. 255.

Miller, D. M., 1980, Structural Geology of the northern Albion Mountains, south-central Idaho., *in* Cordilleran Metamorphic Core Complexes, Crittenden, M. D., Coney, P. J., and Davis, G. H., eds., Geological Society of America Memoir 153, p. 399-423.

Miller, D. M., and Bradfish, L. J., 1980, An inner Cordilleran belt of muscovite-bearing plutons.: Geology, v. 8, p. 412-416.

Miller, E. L., Gans, P.B., Wright, J. E., and Sutter, J. F., in press, 1988, Metamorphic history of the east-central Basin and Range province: tectonic setting and relationship to magmatism., *in* Rubey Volume 7, Metamorphism and Crustal Evolution of the Western United States., Ernst, W. G., ed., University of California, Los Angeles, Calif.

Miller, F. K., and Engels, J. C., 1975, Distribution and trends of discordant ages of the plutonic rocks of northeastern Washington and northern Idaho.: Geological Society of America Bulletin, v. 86, p. 517-528.

Monger, J. W. H., 1985, Structural evolution of the southwestern Intermontane Belt, Ashcroft and Hope map areas, British Columbia.: Current Research Part 1A, Geological Survey of Canada Special Paper 85-1A, p. 358-439.

Monger, J. W. H., Clowes, R. M., Price, R. A., Simony, P. S., Riddihough, R. P., and Woodsworth, G. J., 1985, Continent-Ocean transect B2; Juan de Fuca plate to Alberta plains.: Transects Program, Geological Society of America.

Moore, W. J., and McKee, E. H., 1983, Phanerozoic magmatism and mineralization in the Tooele 1^0 x 2^0 quadrangle, Utah., *in* Tectonic and Stratigraphic Studies in the Eastern great Basin, Miller, D. M., Todd, V. R., and Howard, K. A., eds., Geological Society of America Memoir 157, p. 183-190.

Mortensen, J. K., Montgomery, J. R., and Fillipone, J., 1987, U-Pb zircon, monzonite and sphene ages for granitic orthogneiss of the Barkerville terrane, east-central British Columbia.: Canadian Journal of Earth Science, v. 24, p. 1261-1266.

Mountjoy, E. W., and Forest, R., 1986, Revised structural interpretation, Selwyn Range between Ptarmigan and Hugh Allan Creeks, British Columbia - an antiformal stack of thrusts.: Current Research Part A, Geological Survey of Canada Paper 86-1A, p. 177-183.

Mudge, M. R., and Earhart, R. L., 1980, The Lewis thrust fault in the disturbed belt, northwestern Montana.: United States Geological Survey Professional Paper 1174, 18 p.

Murphy, D. C., 1985, Kinematic and tectonic implications of the vertical transition from upright to near-recumbent folds, central Cariboo Mountains, British Columbia.: Geological Society of America Abstracts with Programs, v. 17, p. 372.

Nichols, D. J., Perry, W. J. JR., and Haley, J. C., 1985, Reinterpretation of the palynology and age of Laramide syntectonic deposits, southwestern Montana, and revision of the Beaverhead Group.: Geology, v. 13, p. 149-153.

Okulitch, A. V., 1985, Assessment of evidence for Paleogene crustal extension in the southern Shuswap metamorphic complex (SSMC), B.C.: Geological Society of America Abstracts with Programs, v. 17, p. 398.

Orr, K. E., 1985, Structural features along the margin of the Okanogan Dome, Tenas Mary Creek area, NE Washington.: Geological Society of America Abstracts with Programs, v. 17, p. 398.

Parkinson, D., 1985, Geochronology of the western side of the Okanogan metamorphic core complex, southern British Columbia.: Geological Society of America Abstracts with Programs, v. 17, p. 399.

Parrish, R. R., 1981, Geology of the Nemo Lakes belt, northern Valhalla Range, southeast British Columbia.: Canadian Journal of Earth Science, v. 18, p. 944-958.

Parrish, R. P., and Wheeler, J. O., 1983, A U-Pb zircon age from the Kuskanax batholith, southeastern British Columbia.: Canadian Journal of Earth Science, v. 20, p. 1751-1756.

Pell, J., and Simony, P. S., 1984, Stratigraphy of the Hadrynian Horsethief Creek Group between the Azure and North Thompson Rivers, Cariboo Mountains, British Columbia.: Current Research Part A, Geological Survey of Canada Paper 84-1A, p. 95-98.

Perry, W. J., Jr., and Sando, W. J., 1982, Sequence of deformation of Cordilleran thrust belt in Lima, Montana region., *in* Geologic studies in the Cordilleran fold and thrust belt, Powers, R. B., ed., Rocky Mountain Association of Geologists, p. 137-144.

Perry, W. J., Jr., Wardlaw, B. R., Bostick, N. H., and Maughan, E. K., 1983, Structure, burial history, and petroleum potential of frontal thrust belt and adjacent foreland, southwest Montana.: American Association of Petroleum Geologists Bulletin, v. 67, p. 725-743.

Potter, C.J., Sanford, W. E., Yoos, T. R., Prussen, E. I., Keach, R. W., II, Oliver, J. E., Kaufman, S., Brown, L. D., and 1986, , Cocorp deep seismic reflection traverse of the interior of the North American Cordillera, Washington and Idaho: Implications for orogenic evolution.: Tectonics, v. 5, p. 1007-1025.

Price, R. A., 1981, The Canadian foreland thrust and fold belt in the southern Canadian Rocky Mountains., *in* Thrust and Nappe Tectonics, McClay, K. R., and Price, N. J., eds., Geological Society of London Special Publication 9, p. 427-448.

Price, R. A., and Mountjoy, E. W., 1970, Geologic structure of the Canadian Rocky Mountains between Bow and Athabasca rivers - a progress report., *in* Structure of the southern Canadian cordillera, Wheeler, J. O., ed., Geological Association of Canada Special Paper 6, p. 7-25.

Read, P. B., and Brown, R. L., 1981, Columbia River Fault Zone: southeastern margin of the Shuswap and Monashee complexes, southern British Columbia.: Canadian Journal of Earth Science, v. 18, p. 1127-1145.

Rhodes, B. P., and Hyndman, D. W., 1984, Kinematics of mylonites in the Priest River "metamorphic core complex", northern Idaho and northeastern Washington.: Canadian Journal of Earth Science, v. 21, p. 1161-1170.

Royse, F., 1985, Geometry and timing of the Darby-Prospect-Hogsback thrust fault system, Wyoming.: Geological Society of America Abstracts with Programs, v. 17, p. 263.

Royse, F., Jr., Warner, M. A., and Reese, D. L., 1975, Thrust belt structural geometry and related stratigraphic problems, Wyoming-Idaho-northern Utah., *in* Symposium on deep drilling frontiers in the central Rocky Mountains, Bolyard, D. W., ed., Rocky Mountain Association of Geologists, p. 41-55.

Ruppel, E. T., 1978, Medicine Lodge thrust system, east-central Idaho and southwest Montana.: United States Geological Survey Professional Paper 1031, 23 p.

—— 1985, Lemhi Arch: a late Proterozoic and early Paleozoic landmass, Central Idaho.: American Association of Petroleum Geologists Bulletin, v. 69, p. 864.

Ruppel, E. T., and Lopez, D. A., 1984, The thrust belt in southwestern Montana and east-central Idaho.: United States Geological Survey Professional Paper 1278, 41 p.

Schirmer, T. W., 1985, Structural evolution of a duplex beneath the Willard thrust, Wasatch Range, Odgen, Utah.: Geological Society of America Abstracts with Programs, v. 17, p. 264.

Silberling, N. J., and Roberts, R. J., 1962, Pre-Tertiary stratigraphy and structure of northwestern Nevada.: Geological Society of America Special Paper 72, 58 p.

Silberman, M. L., and McKee, E. H., 1971, K-Ar ages of granitic plutons in north-central Nevada: Isochron/West, v. 1, p. 15-32.

Skipp, B., 1985, Cabin thrust plate in the Beaverhead Mountains, Idaho and Montana.: Geological Society of America Abstracts with Programs, v. 17, p. 265.

Snoke, A, W, 1980, Transition from infrastructure to superstructure in the northern Ruby Mountains, Nevada., *in* Cordilleran Metamorphic Core Complexes, Crittenden, M. D., Coney, P. J., and Davies, G. H., eds., Geological Society of America Memoir 153, p. 287-333.

Snoke, A. W., and Miller, D. M., in press, 1988, Metamorphic and tectonic history of the northeastern Great Basin., *in* Rubey Vol. 7, Metamorphism and Crustal Evolution of the Western United States., Ernst, G., ed., University of California, Los Angeles, Calif.

Speed, R.C., 1979, Collided Paleozoic microplate in the western United States.: Jour. Geology, v. 87, p. 237-252.

Speed, R. C., Elison, M. W., and Heck, F. R., in press, 1988, Phanerozoic tectonic evolution of the Great Basin., *in* Rubey Volume 7, Metamorphism and Crustal Evolution of the Western United States, Ernst, G., ed., University of California, Los Angeles, Calif.

Speed, R. C., and Sleep, N. H., 1982, Antler orogeny and foreland basin: A model.: Geological Society of America Bulletin, v. 93, p. 815-828.

Sutter, J. F., Snee, L. W., and Lund, K., 1984, Metamorphic, plutonic, and uplift history of a continental-island arc suture zone, west-central Idaho.: Geological Society of America Abstracts with Programs, v. 16, p. 670-671.

Thompson, R. I., 1979, A structural interpretation across part of the northern Rocky Mountains, British Columbia, Canada.: Canadian Journal of Earth Science, v. 16, p. 1228-1241.

Wells, R. E., and Heller, P. L., 1988, The relative contribution of accretion, shear, and extension to Cenozoic tectonic rotation in the Pacific Northwest.: Geological Society of America Bulletin, v. 100, p. 325-338.

Wiltschko, D. V., and Dorr, J. A., Jr., 1983, Timing of deformation in overthrust belt and foreland of Idaho, Wyoming and Utah.: American Association of Petroleum Geologists Bulletin, v. 67, p. 1304-1322.

Winston, D., and Woods, M., 1986, Road Log 3; A travers across the northern Belt basin from East Glacier Park, Montana to Bonners Ferry, Idaho., *in* Belt supergroup: a guide to Proterozoic rocks of western Montana and adjacent areas., Roberts, S. M., ed., Montana Bureau of Mines and Geology, p. 47-68.

Woodward, L. E., 1981, Tectonic framework of disturbed belt of west-central Montana.: American Association of Petroleum Geologists Bulletin, v. 65, p. 291-302.

—— 1982, Tectonic map of the fold and thrust belt and adjacent areas, west-central Montana.: Montana Bureau of Mines and Geology Geologic Map 30, scale 1:250,000.

Woodward, N. B., 1986, Thrust fault geometry of the Snake River Range, Idaho and Wyoming.: Geological Society of America Bulletin, v. 97, p. 178-193.

TABLE 1. Displacement magnitude, period and rate for various positions along the fold and thrust belt.

Location (see Fig. 1)	Horizontal Displacement	Reference	Displacement Onset	Reference	Displacement End	Reference	Displacement period	Displacement rate
A	$120km_a$	Brown et al, 1986; Campbell et al, 1982	$152\pm7_1$	Thompson, 1979	66_2	Thompson, 1979	86my	$1.40\pm.37$mm/yr
B	$200km_a$	Price, 1981; Price and Mountjoy, 1970; Brown et al, 1986	$152\pm7_1$	Price, 1981	$40\pm2_3$ $58\pm2_2$	Price,1981 "	112my	$2.13\pm.81$mm/yr
C	$170\text{–}180km_a$	Harrison et al, 1980; Bally, 1984	$115\pm3_1$ $154\pm2_4$	DeCelles,1986 "	$49\pm9_3$ 56_5 46.3_6 58.3_6	Bally, 1984 Hoffman et al, 1976 Mudge and Earhart, 1980	66my	2.65 ± 1.18mm/yr
D	$150\text{–}180_{b,c}$	Chesson et al, 1984; Hyndman, 1979; Chase et al, 1983; Lagenson et al, 1984a,b	$115\pm3_1$ $154\pm2_4$	DeCelles, 1986 "	56_2 $57.7\pm1.4_6$	DeCelles et al, 1987 "	59my	2.8 ± 1.3mm/yr
E	$140\text{–}150km_a$	Jordon, 1981; Royse et al, 1975	$145\pm5_1$	Armstrong and Cressman, 1983	$52\pm2_3$	Wiltschko and Dorr, 1983; Royse et al, 1975	93my	$1.56\pm.12$mm/yr
			*116 ± 3	Heller et al, 1986			62my	$2.34\pm.94$mm/yr
F	$100\text{–}120km_a$	Allmendinger et al, 1986	$105\pm7_1$	Villien and Kligfield, 1986	$55\pm3_3$	Villien and Kligfield,1986	50my	$2.2\pm.56$mm/yr

TABLE 1. (cont)

a) displacement magnitude by palinspastic reconstruction

b) displacement magnitude by summation of individual thrust displacements

c) displacement magnitude by correlation to north and/or south

1) displacement onset by earliest foreland basin sediments

2) displacement end by cessation of foreland basin sedimentation

3) displacement end by overlapping strata

4) possible displacement onset by youngest strata beneath foreland basin sediments

5) displacement end by youngest burial metamorphism - radiometric age

6) displacement end by date of cross-cutting intrusion - radiometric age

TABLE 1. (continued)

Thrust	Displacement	Period	Rate	Reference
Lewis	65km	16my (72 - 56)	4.06mm/yr	Mudge and Earhart, 1980 Hoffman and others, 1976
Lewis - Defo. Front	90km	30my (70 - 40?)	3.0mm/yr	Price, 1981
McConnel - Defo. Front	75km	30my	2.5mm/yr	Price, 1981
Paris-Willard	18km	33my (145 - 112)	0.55mm/yr	Allmendinger, in press
	40km	48my (140 - 92)	0.83mm/yr	Wiltshko and Dorr, 1983
Crawford-Meade	50km	16my (106 - 90)	3.13mm/yr	Allmendinger, in press
	29km	43my (90 - 47)	0.67mm/yr	Wiltshko and Dorr, 1983
Absaroka (system)	40km	10my (80 - 70)	4.0mm/yr	Allmendinger, in press
	14km	18my (80 - 62)	0.77mm/yr	Wiltshko and Dorr, 1983
	42km	----	----	Woodward, 1986
	15 - 48km	----	----	Dixon, 1982
Darby (system)	40km	13my (68 - 55)	3.07mm/yr	Allmendinger, in press
	10km	8my (64 - 56)	1.25mm/yr	Wiltshko and Dorr, 1983
	29km	----	----	Royce, 1985
	16 - 41km	----	----	Dixon, 1982

TABLE 2. DISPLACEMENT RATE ESTIMATES FOR THE CONTINENTAL HINTERLAND.

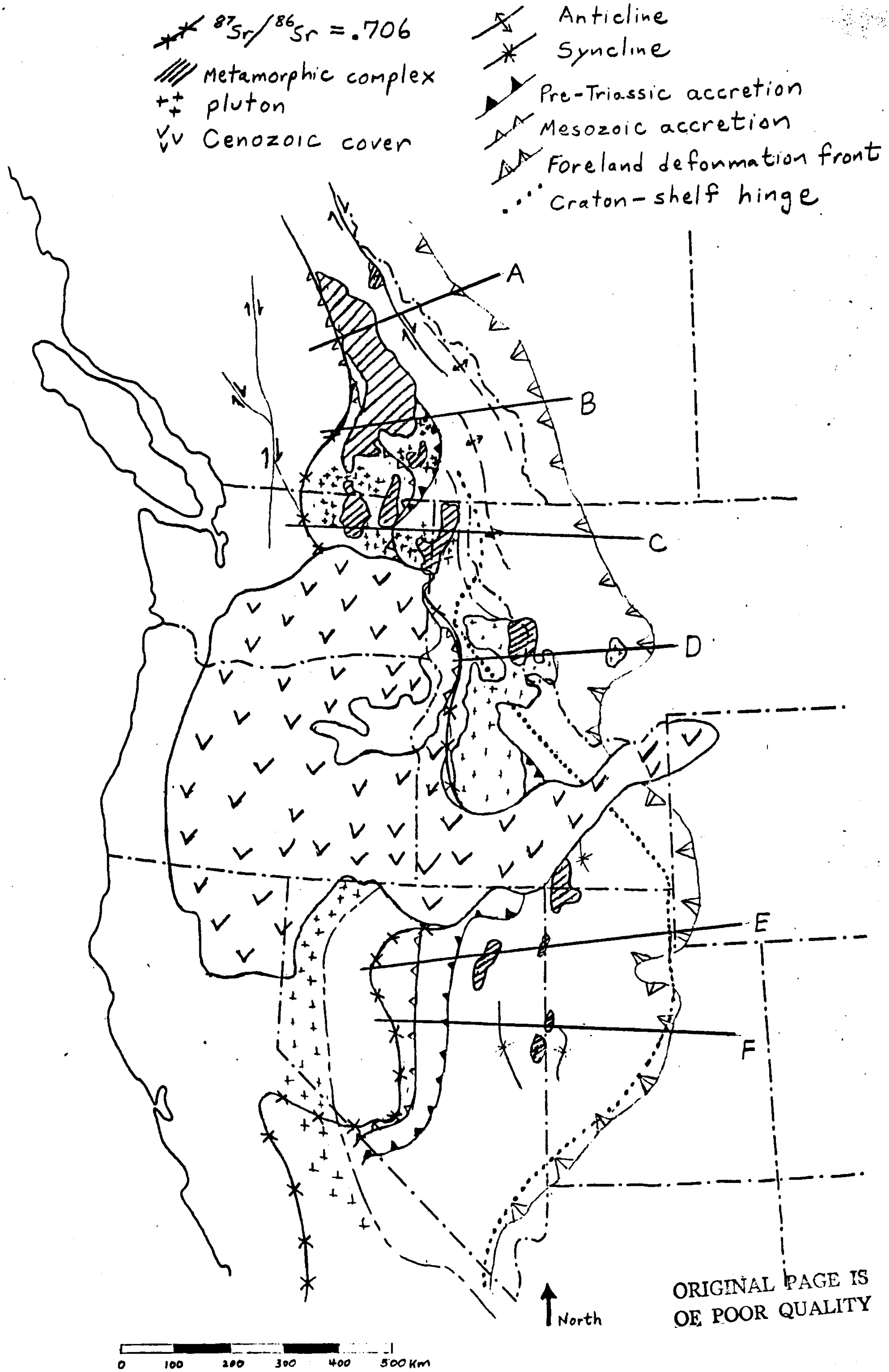
Location	Vertical Displacement	Displacement Period	Displacement Rate	Reference
Monashee Decollement	10-12km	75my (155/165) - (80/90)	0.15 mm/yr	Journeay and Brown, 1986
Kootenay Arc	20km	10my (165 - 155)	2.0 mm/yr	Archibald and others, 1983
Okanogan Dome	10km	10my (96 - 55)	1.0 mm/yr	Parkinson, 1985
Idaho Batholith	20km?	35my (90/70) - (40/50)	0.57 mm/yr	Hyndman, 1979 Chase and others, 1983
Ruby Range	13km	25my (45 - 20)	0.5 mm/yr	Dallmeyer and others, 1986
Ruby Range	6.5km	115my (160 - 45)	0.006 mm/yr	Dallmeyer and others,]1986

TABLE 3. PREDICTED CRUSTAL THICKENING OF CONTINENTAL HINTERLAND

Position	Current Width	Extension	Width(pre-extension)	Foreland Shortening	Thickening
A	125km	0%	125km	120km	1.96
B	200km	20%	160km	200km	2.25
C	325km	40%	195km	180km	1.92
D	100km?	----	----	150km?	2.5??
E	550km	100%	225km	140km	1.62
F	550km	100%	225km	110km	1.49

Figure Captions

Figure 1. Map of the western North American Cordillera showing structural and thermal features discussed in text. Lines with corresponding letters A through F are the locations for which displacement magnitude, period, and rate data are presented.



$^{87}Sr/^{86}Sr = .706$
Metamorphic complex
pluton
Cenozoic cover
Anticline
Syncline
Pre-Triassic accretion
Mesozoic accretion
Foreland defonmation front
Craton-shelf hinge
A
B
C
D
E
F
North
0 100 200 300 400 500 Km

FIG. 1

APPENDIX 3

ALLEGHANIAN DISPLACEMENTS IN THE APPALACHIAN MOUNTAINS

Frederick R. Heck

Department of Geological Sciences

Northwestern University

Introduction

This paper is a synthesis of the displacement history of rocks in the Appalachian mountains that were transported during the late Paleozoic Alleghanian orogeny. The Alleghanian is the youngest of three major tectonothermal events (Taconic, Acadian, Alleghanian) that built the Appalachian Mountains during the Paleozoic Era. The goals of this synthesis are to determine how the magnitudes, directions, and rates of Alleghanian displacements vary with position in the mountain belt and to make interpretations regarding the causes of such variation. The history of Alleghanian displacements is also required to reconstruct the pre-Alleghanian position of lithotectonic assemblages, an essential first step in reconstructing the older orogenic events.

A striking feature of Alleghanian deformation is the position in the orogen of strike-slip deformation. Strike-slip displacements parallel to the mountain belt are confined to regions outboard of the buried North American craton edge where they occur with compressional displacements normal to the orogen as part of a transpressional deformation. Inboard of the craton edge only orogen-normal compressional displacements occur. This region of strike-slip displacements is also where Alleghanian high-grade metamorphism and plutonism occur in a belt that extends from Georgia to Pennsylvania. We argue that the craton edge and belt of high-grade metamorphism and plutonism exerted fundamental controls on the position of orogen-parallel displacements during the Alleghanian. Such relationships may provide criteria for recognizing such features in other mountain belts, ancient or modern.

Timing of Appalachian tectonic events

The tectonic history of the Appalachians begins in the late Precambrian and Early Cambrian with formation of an east-facing passive margin by rifting and spreading between the North American craton and terranes to the east (Rast and Kohles, 1986). Closure of the resulting ocean basin(s) occurred through the Paleozoic and gave rise to three major tectonothermal events that built the Appalachian Mountains (Hatcher, 1987; Glover and others, 1983). The Taconic orogeny occurred in Middle and Late Ordovician (~480 - 430 Ma; Zen, 1983), the Acadian occurred in Early through Late Devonian (~400 - 360 Ma; Zen, 1983) in the northern Appalachians and extended into Mississippian time in the southern Appalachians (~380 - 340 Ma; Glover and others, 1983). The Alleghanian marked the final closure of the early Paleozoic ocean basin(s) with collision of northwest Africa and east margin of North America (Arthaud and Matte, 1977; Secor and others, 1986a; Vauchez and others, 1987). Estimates of the Alleghanian time interval vary from ~300 - 250 Ma (Zen, 1983) to ~330 - 230 Ma (Glover and others, 1983). We prefer the longer time interval because: 1) the base of the Alleghanian clastic wedge in the Valley and Ridge province occurs in rocks of Late Mississippian age (333 +/-22 to 320 +/-21 Ma according to Palmer (1983)), 2) a well defined episode of granitoid pluton emplacement occurred in the eastern Piedmont province of the southern and central Appalachians between about 330 and 285 Ma (Sinha and Zietz, 1982) and similar plutons with ages of 325 to 275 Ma occur as far north as central Maine (Aleinikoff and others, 1985), 3) the Alleghanian metamorphic peak in the southern Appalachians was 315 +/-24 to 295 +/-5 Ma (Dallmeyer and others, 1986), 4) metamorphic cooling ages probably related to uplift during broad folding are at least as young as 238 Ma in the southern Appalachians (Russell and others, 1985), and 5) compressional deformation followed a cooling age of 243 +/-5 Ma in the Honey Hill fault zone in southern Connecticut (Wintsch and Sutter, 1986). The Paleozoic orogenic events were followed by Late Triassic to Jurassic rifting and spreading during which time extensional structures overprinted and often reactivated Paleozoic structures.

Alleghanian deformation

The type area of Alleghanian deformation is the Valley and Ridge province of the southern and central Appalachians (Fig. 1) where Cambrian to Permian shallow marine strata are deformed in a foreland fold and thrust belt. Significant Alleghanian deformation also occurred in the eastern Piedmont of the southern and central Appalachians from Alabama to Pennsylvania. Deep levels of Alleghanian deformation are exposed in the eastern Piedmont (>10 - 15 km) whereas Valley and Ridge deformation occurred at shallow levels. The Blue Ridge-Inner Piedmont, which was strongly deformed and metamorphosed in the Taconic (less so in the

Acadian), mostly escaped Alleghanian deformation and was transported as a semirigid block that overthrust and shortened previously undeformed shallow marine strata of the Valley and Ridge (Glover and others, 1983; Hatcher, 1987). In the northern Appalachians, where most deformation is Taconic and Acadian, Alleghanian effects are mostly confined to the eastern side of the orogen; no Valley and Ridge type fold and thrust belt is developed in the west. Deep levels of Alleghanian deformation are exposed as far north as southern New England and shallower levels are exposed to the north in Maine and the Canadian maritime provinces.

Figures 1b and 2 show the position and timing of Alleghanian structures and deformation phases. The northern Appalachians are characterized by dextral transpression in the Alleghanian which, at least in New Brunswick and Nova Scotia, also characterizes the transition from Acadian to Alleghanian time but with a somewhat different shortening direction. In the Acadian-Alleghanian transition dextral strike-slip occurred on north to northeast trending faults (Webb, 1969) indicating southwest to west maximum compression. In the Alleghanian, dextral strike-slip occurred on east trending major faults (Cobequid-Chedabucto, offshore extension of the Bloody Bluff fault, and probably the Martic zone) (Eisbacher, 1970; Webb, 1969; Hill, 1986; Simpson and others, 1980; Song and Hill, 1988) that transfer displacements at their west ends to northwest or southeast verging folds and thrusts (St. John fold and thrust belt, Bloody Bluff-Honey Hill faults, northern Valley and Ridge province) (Nance, 1986; Smith and Barosh, 1983; Hepburn and others, 1987; Geiser and Engelder, 1983) indicating a change to west-northwest - east-southeast maximum compression. The transpressive regime produced the Cobequid uplift in Nova Scotia, probably produced a postulated highland (Rast, 1984) along the offshore extension of the Bloody Bluff fault, and may be responsible for uplift north of the Martic zone that is indicated by a northeast to southwest increase in Alleghanian metamorphic grade (O'Hara, 1986) and exposed depth of Alleghanian deformation that occurs from Maine to southern Connecticut and Rhode Island (Ludman, 1986; Wintsch and Sutter, 1986). In the Late Permian into Triassic time maximum compression directions on the Honey Hill fault rotated clockwise from west-northwest - east-southeast to northwest - southeast to north - south (Wintsch and Sutter, 1986). Similar clockwise rotations occurred at about the same time in the Valley and Ridge in Pennsylvania (Nickelsen, 1979; Geiser and Engelder, 1983) and in the Green Pond outlier in the New Jersey Highlands (Mitchell and Forsyth, 1988).

Dextral transpression during west to northwest maximum shortening also characterizes Alleghanian deformation in the Piedmont province of the southern and central Appalachians where Alleghanian plutonism and high-grade metamorphism occur in a southwest-northeast trending belt extending from Alabama to Pennsylvania. At each of several well-studied localities from Virginia to Alabama the boundary-normal component of transpression produced one to two generations of northeasterly trending major or minor folds with cleavage at conditions of upper greenschist to amphibolite grade metamorphism. In the eastern Piedmont, orogen-parallel dextral strike-slip on northeast trending mylonite zones (eg., the Hylas, Nutbush Creek, Hollister, Modoc, Towaliga, and Goats Rock zones) mostly follows the folding and is often localized along the limbs of earlier major folds (Bobyarchick and Glover, 1979; Farrar, 1985; Secor and others, 1986b). Farther west, near the western limit of strike-slip displacement at the Inner Piedmont-Blue Ridge boundary, strike-slip displacements are older and precede or are synchronous with boundary normal compression (Brevard zone, Bowens Creek zone, Brookneal zone) (Gates and others, 1986; Gates, 1987; Edelman and others, 1987; Vauchez, 1987; Steltenpohl and others, 1988). Alleghanian deformation also occurs along steep mylonite zones in the Kings Mountain Belt (Horton and others, 1987). Although the kinematics of these zones is not worked out, their age and orientation suggest they too experienced dextral-transpressive displacements.

Alleghanian deformation in the Valley and Ridge province is due to west to northwest compression with no component of orogen parallel strike-slip displacement. The Valley and Ridge province is a foreland fold and thrust belt in which displacements propagated westward above a regional detachment that overlies the craton and extends east beneath the Inner Piedmont-Blue Ridge thrust sheet to the Kings Mountain Belt (Cook, 1979; Boyer and Elliot, 1982). The southern Valley and Ridge is characterized by numerous thrust sheets whereas in the central Valley and Ridge few thrusts break the surface although they are known to core anticlines and merge at depth in a regional detachment (Kulander and Dean, 1986. This change in style reflects greater shortening in the southern Valley and Ridge. There is also a change in structural trend from north-northeast in the central Appalachians to northeast in the southern Appalachians. The trend change occurs in the Roanoke recess where maximum compression directions rotate counter-clockwise from N10-30W to N50-70W to N85W (Dean and others, 1988). In the southernmost Appalachians of Alabama there is a clockwise rotation of maximum compression directions from west-northwest - east-southeast to northwest-southeast to north-south (Tull, 1984).

Magnitudes, directions, and rates of Alleghanian displacements

Magnitudes of Alleghanian displacement due to northwest-southeast shortening are best known in the Valley and Ridge fold and thrust belt where magnitudes are determined from balanced cross-sections and/or deformed versus undeformed line lengths. Displacement directions are assumed to be parallel to the cross-sections that trend about normal to strike. The magnitudes and directions are given with respect to undeformed North American continent. For example, the Blacksburg, Virginia cross-section (section 3, Fig. 1b) estimates Valley and Ridge shortening between the undeformed continent and the trailing edge of the Pulaski thrust sheet at a point just west of the Blue Ridge-Inner Piedmont thrust sheet. The estimated shortening is the sum of the shortening within the Pulaski sheet (ranging from 80 to 136 km with an average of 108 km; Bartholomew, 1987) plus the displacement along the Pulaski thrust (100 to 110 km, average 105 km; Bartholomew, 1987) plus the shortening due to folds and thrusts below and west of the Pulaski sheet (33 to 59 km, average 46 km; Kulander and Dean , 1986). The best pick total shortening is therfore 259 km although the amount could vary by as much as 46 km. The displacement direction is taken to be N28W parallel to the cross-section trend which is normal to structural trends at Blacksburg. Although northwest-southeast Alleghanian shortening clearly occurred in the high metamorphic grade eastern Piedmont, no shortening estimates are available because suitable marker horizons are absent and Alleghanian effects are difficult to distinguish from older deformations except where radiometric ages are available, and in those cases shortening estimates have not been done. The same situation applies to the northern Appalachians.

Alleghanian strike-slip displacements occur only in the Piedmont province in the southern and central Appalachians and in the easternmost portions of the northern Appalachians. Magnitudes of strike-slip displacement are determined by the offset of features across the strike-slip zone (eg., plutons, isopach contours, terrane boundaries) or by estimates of simple shear strain versus distance across the zone. Displacement directions are assumed to be parallel to the fault zone or to mineral elongation iineations if they are present. In all cases the magnitudes of strike-slip displacement are more imprecise and poorly constrained than are displacements in the Valley and Ridge because: 1) area balancing is not possible as in Valley and Ridge cross-sections, 2) the types of offset features have boundaries that are generally unsuitable for precise measurements of offset, 3) unrecognized dip-slip motions may be present which could result in strike-slip offsets that are apparent and not real, and

4) displacement estimates by shear strain versus distance across the shear zone invariably assume simple shear alone has occurred with no component of shortening normal to the zone (an invalid assumption in most cases) and the position of shear zone boundaries is often poorly constrained.

Magnitudes and directions of strike-slip displacements are given with respect to a point on the opposite side of the strike-slip zone rather than to undeformed North America. To find the displacement with respect to undeformed North America the displacement vector for the fault in question must be added to all orogen-parallel and orogen-normal displacement vectors between the fault and undeformed North America (Fig. 3). However, the uncertainties in magnitude and direction of such a vector are probably quite large because the procedure assumes that all displacement vectors between undeformed North America and the point in question are known. The assumption is acceptable for Valley and Ridge displacements but clearly not for the Piedmont or the northern Appalachians where an unknown amount of northwest shortening occurred in the Alleghanian and where all zones of strike-slip displacement have probably not been identified or if identified their displacements are not always known (eg., the Brevard zone, and others). Therefore, at this time we do not present the strike-slip displacements with respect to a point on undeformed North America although this clearly is an eventual goal of the study.

As an example of strike-slip displacement estimates, the Brookneal shear zone near Brookneal, Virginia (Gates and others, 1986) is about 4 km wide with a shear plane orientation of N40E, 50SE. Shear zone fabrics indicate dextral displacement subparallel to the zone. Displacement magnitude is estimated as > 17km by determining simple shear strain across the zone according to changing cleavage orientations. The displacement magnitude is considered to be a minimum because simple shear is heterogeneous across the zone and all areas of high simple shear have probably not been recognized. Furthermore, the 4 km width of the zone is probably a minimum value. If the zone is transpressive, however, a component of shortening normal to the shear plane will produce the observed cleavage relations at lower simple shear strains which will in turn reduce the estimated strike-slip displacement. Elongate mineral lineations plunge shallowly N50E which when considered with the shear plane orientation and dextral sense of shear indicates a small component of up dip displacement. Furthermore, the Bowen Creek fault zone to the northwest is related to the Brookneal zone and is clearly transpressive (Gates, 1987). Therefore, the Brookneal zone probably experienced at least some component of shortening normal to the zone which has the effect of reducing the minimum displacement estimate (although we have no way of knowing by how much). The displacement direction of rocks outboard of the Brookneal zone is taken as S50W; dextral displacement parallel to the shear zone stretching lineations.

Rates of Alleghanian displacement are more uncertain than magnitudes and directions due to the added uncertainty in the timing of displacement. Displacement timing is reasonably well constrained for individual structures in the Piedmont province and northern Appalachians because cross-cutting intrusives, metamorphic ages, and overlapping strata are common. In the Valley and Ridge however, such features do not occur; timing of deformation cannot be evaluated independently at different positions so the province must be treated as a single package with respect to time interval of deformation. Good estimates of the time of Valley and Ridge deformation are provided by recently published K/AR illitization ages (303 +/-13 Ma to 265 +/-13 Ma) from the Valley and Ridge and Plateau provinces of Alabama, Tennessee, Kentucky, and Virginia (Elliot and Aronson, 1987). These ages are compatible with stratigraphic constraints on timing of deformation. They follow and partly overlap the ages of early Alleghanian clastic wedge strata that were deposited in the Valley and Ridge during uplift to the east and later deformed as Alleghanian displacements migrated westward. Also, the age of the youngest preserved unit affected by Alleghanian deformation, the Dunkard Group (286 +/-12 Ma to 266 +/-17 Ma; Secor and others, 1986a), falls within the time interval of illitization. Deformation of the Dunkard probably represents the waning stages of Alleghanian deformation because it is among the westernmost of deformed strata and is weakly deformed in broad, open, upright folds. Alleghanian deformation therefore probably ended shortly after this interval. A more conservative estimate of the end of Alleghanian deformation in the Valley and Ridge is 235 Ma which is about the youngest age of Alleghanian deformation recognized in the Appalachians (see above). We therfore use 38 Ma (303 - 265 Ma) as a minimum time interval and 81 Ma (316 - 235 Ma) as a maximum time interval of Alleghanian displacement in the Valley and Ridge province.

Table 1 presents displacement rates of Alleghanian strike-slip and fold/thrust structures for which the required displacement magnitude and timing data are available. The displacement rates were determined in the following manner. For each point we choose two values of displacement magnitude and two values of time interval over which displacement occurred. The two displacement magnitude values are the maximum and minimum estimates in the cited reference. If only a single value of displacement was provided in the cited reference we use 25% less and 25% more of the value as minimum and maximum estimates of displacement. The two time interval values are the interval which is the best pick estimate in the cited reference and the interval which equals the best pick plus the uncertainties at each end of the best pick. For example, if the time of displacement is given as between 270 +/-15 Ma and 250 +/-10 Ma the two time interval values we use are 20 Ma (270 - 250 Ma) and 45 Ma (285 - 240 Ma) (see below for how we use them). If the cited reference does not provide uncertainties at each end of the best pick time interval, we add 50% of the best pick as a measure of uncertainty to obtain the second time interval. Where timing of displacement is given by the age of stratal intervals (eg., Pennsylvanian age) we use the DNAG best pick ages (Palmer, 1983) for our best pick interval (320 - 286 = 34 Ma interval for the Pennsylvanian) and add in the uncertainties of the DNAG best picks for a maximum estimate of the displacement time interval (34 + 20 + 12 = 66 Ma). In some cases deformation is estimated to begin between two times and end between two other times. For example, deformation began between 290-280 Ma and ended between 260-250 Ma. In this case a minimum time interval is 20 Ma (280 - 260 Ma) and a maximum interval is 40 Ma (290 - 250). These values are used with the maximum and minimum displacement magnitudes respectively to determine maximum and minimum displacement rates. The best pick time interval for calculating displacement rate is taken as the avervage time interval or 30 Ma.

The best pick displacement rates in Table 1 are determined by dividing the average displacement by the best pick displacement time interval. In the case where displacement occurred within the best pick interval the uncertainties are given as the difference between the best pick rate and minimum rate which is the minimum displacement magnitude divided by the maximum time interval (equal to the best pick time interval plus the uncertainties at each end of the best pick interval). For example, assume the displacement magnitude is 60 to 80 km and the time interval of displacement is between 270 +/-15 Ma and 250 +/-10 Ma. The best pick displacement rate is 70km / 20Ma or 3.5 mm/yr. The minimum rate is 60km / 45Ma or 1.33 mm/yr. The uncertainty is therfore 3.5 - 1.33 = 2.17 and the displacement rate is displayed as 3.5 (+/- 2.17) mm/yr. This procedure is adopted because in some cases, as in this example, a minimum time interval (which with the maximum displacement magnitude would provide an independent maximum rate) obtained by subtracting the uncertainties from the best pick interval gives a negative time interval of displacement. In the case where displacement is constrained to begin within one time interval and end within another the uncertainties are given as the difference between the best pick rate and the minimum and maximum rates.

In almost all cases the best pick displacement rates represent minimum values because the displacement time interval represents the time within which displacement occurred. For example, a fault offsets a pluton dated at 270 Ma and is pinned by another pluton dated at 240 Ma. Displacement on the fault occurred over some time interval between 270 and 240 Ma. The true rate is therefore greater than the calculated rate which considers the entire time interval. The effect is exaggerated because displacement estimates are commonly minimum values due to the presence of cleavage and other small scale structures whose shortening effects are not considered in the displacement estimate.

The major problems in determination and interpretation of the Alleghanian displacement rate data are: 1) the use of the same time interval for displacements all along the Valley and Ridge, and 2) the relatively poorly constrained displacement magnitudes from the Piedmont province. Because displacement timing cannot be determined independently for each cross-section in the Valley and Ridge along which displacement magnitudes are determined, we must use the same timing data for each cross-section. The time interval of displacement is therefore a constant and displacement rates vary directly with displacement magnitude. Our calculated rates are therefore greater in the southern than in the central Appalachians because the displacement magnitudes are greater in the south. Rates may indeed be greater in the south but it may also be true that deformation there occurred over a longer time interval which would tend to equalize the displacement rates. In contrast to the Valley and Ridge, displacement time intervals in the Piedmont province are relatively well constrained for individual stuctures because of abundant metamorphic/plutonic ages but estimates of displacement magnitude are more uncertain for reasons discussed above.

Position and shape of early Paleozoic craton margin

The position and shape of the early Paleozoic craton margin (Fig. 1b) is interpreted from both geophysical and geological data. In the southern and central Appalachians the margin is marked by a strong gravity gradient that indicates thick continental crust to the west and tectonically thinned, attenuated crust to the east (Hatcher and Zietz, 1980). The gravity gradient coincides with a regional northeast trending monoclinal flexure in which rocks fabrics have shallow dips west of the gradient and steep northwest-southeast dips east of the gradient (Price and Hatcher, 1983). Reflection seismic data tend to support the interpretation by indicating zones of east dipping reflectors proximal to the position of the gravity gradient (Ando and others, 1983; Nelson and others, 1985). The strike-parallel shape of the craton margin in the southern and central Appalachians indicated by the gravity gradient is also reflected in the change in trend of Alleghanian surface structures, especially the east-west trend in southern Pennsylvania. Stratigraphic evidence indicates the shape is inherited from late Precambrian rifting (Thomas, 1977) rather than a result of Paleozoic deformation. The west-southwest - east-northeast trend of the craton edge in southern Pennsylvania probably caused the clockwise rotations of maximum compression axes (see above) during regional west to west-northwest shortening. Rotation of compression axes in Virginia and Alabama are also probably due to shortening against the irregular craton edge.

In the northern Appalachians the gravity data are more ambiguous. However, when considered with surface geologic data (positions of ophiolites, blueschist metamorphic rocks, melange zones, island arc rocks, etc) a reasonable estimate can be made of the position of the craton edge (Zen, 1983). As in the southern and central Appalachians, structural trends are parallel to the craton edge indicating it exerted a fundamental control on deformation patterns, although in the northern Appalachians the deformation thus controlled was mostly Taconic and Acadian in age rather than Alleghanian whose effects were relatively minor in the north.

An important problem of Appalachian tectonics is the degree to which Mesozoic extension has produced the geophysical (and geological?) observations we see today. The marked gravity gradient that defines the craton edge is bounded on the east by a northeast trending belt of gravity highs that give way farther east to lower gravity values. This gravity trough is commonly assumed to be a Paleozoic structure (eg., a paleosubduction zone, Thomas, 1983) but it could mark a belt of deep seated mafic intrusions related to Mesozoic extension. Nelson and others (1986) present arguments that suggest Mesozoic extension may have played a major role in producing present day seismic reflection patterns in the southern and central Appalachians.

Distribution of Alleghanian strike-slip faults

The confinement of Alleghanian strike-slip displacements to the region outboard of the craton margin where the belt of Alleghanian high-grade metamorphism and plutonism occur suggest that one or both of these features may have controlled the position of orogen parallel displacements. The craton margin is a crustal scale boundary that separates thick continental crust to the west from tectonically thinned continental crust or noncontinental crust to the east. Such a feature is a natural anisotropy for controlling the continentward limit of strike-slip deformation. This influence is enhanced by the change in orientation of preexisting rock fabrics that occurs at the margin. Outboard of the craton margin rock fabrics generally have steep northwest-southeast dips due to pre-Alleghanian and early Alleghanian shortening subperpendicular to the margin. Inboard of the margin rock fabrics in the Blue Ridge-Inner Piedmont thrust sheet and in the Valley and Ridge are more flat lying. Rock fabrics outboard of the craton margin were therfore favorably oriented anisotropies for the development of Alleghanian strike-slip faults whereas the absence of such anisotropies inboard of the margin prohibited strike-slip displacements from propagating inboard of the margin. An exception to this is the Brevard zone which experienced dextral slip in the Alleghanian and occurs just inboard of the craton edge. However the Brevard zone was a pre-Alleghanian zone of weakness susceptible to reactivation and it does not extend at depth into the craton but roots in a subhorizontal Alleghanian detachment above the craton. Furthermore, it is possible that the Brevard zone was transported across the craton margin on the Alleghanian detachment after it experienced early Alleghanian dextral strike-slip outboard of the margin.

The coincidence of Alleghanian strike-slip faults with the belt of Alleghanian high-grade metamorphism and plutonism suggests a second possible explanation for the position of strike-slip displacements. In this case the belt of hot ductile rock acted as an orogen-parallel zone of crustal weakness in which strike-slip displacements were accomodated in preference to regions of relatively lower heat flow to the west. This explanation assumes that the metamorphic-plutonic belt existed as a belt in the Alleghanian rather than extending to the west at depth and being exposed today only in the eastern Piedmont. The assumption is probably valid according to arguments by Sinha and Zietz (1982) that this belt is an Alleghanian magmatic arc.

These relationships between the position of Alleghanian strike-slip displacements and the positions of the craton margin and metamorphic-plutonic belt suggest criteria for where to seek such features in other oblique collisional orogens. The foreland-most strike-slip faults may indicate the position of the subsurface craton edge and regions of most abundant strike-slip deformation may correspond to zones of greater thermal activity at depth. Similary, the position of the craton edge and zones of high heat flow may be criteria for investigating the presence of orogen-parallel strike-slip displacements. We are presently investigating whether such criteria apply to active collision zones, or even to other ancient zones.

Conclusions

1. Alleghanian displacement magnitudes in the Appalachian Mountains are reasonably well constrained in the Valley and Ridge province but less well constrained in the Piedmont province and northern Appalachians. Average shortening of the Valley and Ridge, among the sections we have evaluated, is ~218 km with the minimum shortening the north (138 km) and the maximum shortening in the south (269 km). Average shortening direction in the Valley and Ridge is N45W. Among the strike-slip faults evaluated for displacement rate, displacement magnitudes range from 12 to 210 km in the northern Appalachians and >17 to 160 km in the southern Appalachians. The average strike-slip displacement direction for the entire orogen is S52W (S55W in the northern Appalachians and S49W in the central and southern Appalachians.

2. Displacement timing is reasonably well constrained for individual structures in the Piedmont province and northern Appalachians but very poorly constrained for individual structures in the Valley and Ridge. The average best pick time interval of displacement for individual strike-slip structures is 28.5 Ma. This value is comparable to the best pick time interval for Valley and Ridge shortening (38 Ma) which gives some confidence in the more poorly constrained Valley and Ridge value.

3. Valley and ridge displacement rates vary from north to south from ~3.63 to 7.08 mm/yr (5.75 mm/yr average). However, comparison of these rates is of questionable value because the same displacement time interval must be used in the rate calculations.

4. Comparison of strike-slip displacement rates along the eastern Appalachians is probably worthwhile because displacement timing for individual structures is fairly well constrained and in most cases displacement magnitudes are probably correct within their uncertainties. The range and average of strike-slip displacement rates in the northern Appalachians (min. = .74 mm/yr; max. = 5.48 mm/yr; average = 2.79 mm/yr) is similar to those values in the southern Appalachians (min. = >.63 mm/yr; max. = 5.00 mm/yr; average = 2.07 mm/yr).

5. Orogen-parallel displacements are confined to the region outboard of the Early Paleozoic craton margin in the

belt of Alleghanian high-grade metamorphism and plutonism. These relationships may provide criteria for recognizing the preexisting craton edge, deep zones of greater thermal activity, or previously unrecognized zones of strike-slip faulting in other ancient or modern zones of oblique collision.

6. Two important problems of Alleghanian/Appalachian tectonics that require further evaluation are: 1) the amount of Alleghanian age, orogen-normal shortening in the Piedmont province, and 2) the degree to which Mesozoic extension has produced present day geophysical and geological observations that are commonly assumed to be due to Paleozoic tectonics. Both of these problems are part of our ongoing research for NASA.

REFERENCES CITED

Aleinikoff, J.N., Moench, R.H., and Lyons, J.B., 1985, Carboniferous U-Pb age of the Sebago batholith, southwestern Maine: metamorphic and tectonic implications: Geological Society of America Bulletin, v. 96, p. 990-996.

Anderson, F.D., 1972, The Catamaran fault, north-central New Brunswick: Canadian Journal of Earth Science, v. 9, p. 1278-1286.

Ando, C.J., Cook, F.A., Oliver, J.E., Brown, L.D., and Kaufman, S., 1983, Crustal geometry of the Appalachian orogen from seismic reflection studies, *in* Contributions to the Tectonics and Geophysics of Mountain Chains, Hatcher, R.D., Jr., Williams, H., and Zeitz, I., eds., Geological Society of America Memoir 158, p. 83-101.

Arthaud, F., and Matte, P., 1977, Late Paleozoic strike-slip faulting in southern Europe and northern Africa: result of a right-lateral shear zone between the Appalachians and the Urals: Geological Society of America Bulletin, v. 88, p. 1305-1320.

Bartholomew, M.J., 1987, Structural evolution of the Pulaski thrust system, southwestern Virginia: Geological Society of America Bulletin, v. 99, p. 491-510.

Bobyarchick, A.R., and Glover, L., III, 1979, Deformation and metamorphism in the Hylas zone and adjacent parts of the eastern Piedmont in Virginia: Geological Society of America Bulletin, v. 90, p. 739-752.

Boyer, S.E., and Elliot, D., 1982, Thrust systems: American Association of Petroleum Geologists Bulletin, v. 66, p. 1196-1230.

Cook, F.A., Albaugh, D.S., Brown, L.D., Kaufman, S., Oliver, J., and Hatcher, R.D., Jr., 1979, Thin-skinned tectonics in the crystalline southern Appalachians; COCORP seismic-reflection profiling of the Blue Ridge and Piedmont: Geology, v. 7, p. 563-567.

Dallmeyer, R.D., Wright, J.E., Secor, D.T., and Snoke, A.W., 1986, Character of the Alleghanian orogeny in the southern Appalachians: Part II. Geochronological constraints on the tectonothermal evolution of the eastern Piedmont in South Carolina: Geological Society of America Bulletin, v. 97, p. 1329-1344.

Dean, S.L., Kulander, B.R., and Skinner, J.M., 1988, Structural chronology of the Alleghanian Orogeny in southeastern West Virginia: Geological Society of America Bulletin, v. 100, p. 299-310.

Dennis, A.J., and Secor, D.T., 1985, Genesis and geometry of an Alleghanian shear zone in the central South Carolina Piedmont: Geological Society of America Abstracts with Programs, v. 17, p. 87.

Druhan, R.M., Butler, J.R., Horton, J.W., and Stoddard, E.F., 1988, Tectonic significance of the Nutbush Creek fault zone, eastern Piedmont of North Carolina and Virginia: Geological Society of America Abstracts with Programs, v. 20, p. 261.

Edelman, S.H., Liu, A., and Hatcher, R.D., Jr., 1987, The Brevard fault zone in South Carolina and adjacent areas: an Alleghanian dextral shear zone reactivated as a thrust fault: Geological Society of America Abstracts with Programs, v. 19, p. 82.

Eisbacher, G.H., 1970, Deformation mechanics of mylonitic rocks and fractured granites in the Cobequid Mountains, Nova Scotia, Canada: Geological Society of America Bulletin, v. 81, p. 2009-2020.

Elliot, W.C., and Aronson, J.L., 1987, Alleghanian episode of K-bentonite illitization in the southern Appalachian Basin: Geology, v. 15, p. 735-739.

Farrar, S.S., 1985, Tectonic evolution of the easternmost Piedmont, North Carolina: Geological Society of America Bulletin, v. 96, p. 362-380.

Gates, A.E., 1987, Transpressional dome formation in the southwest Virginia Piedmont: American Journal of Science, v. 287, p. 927-949.

Gates, A.E., Simpson, C., and Glover, L., 1986, Appalachian Carboniferous strike-slip faults: an example from Brookneal, Virginia: Tectonics, v. 5, p. 119-133.

Geiser, P., and Engelder, T., 1983, The distribution of layer-parallel shortening fabrics in the Appalachian foreland of New York and Pennsylvania: evidence to two non-coaxial phases of of Alleghanian orogeny, *in* Contributions to the Tectonics and Geophysics of Mountain Chains, Hatcher, R.D., Williams, H., and Zeitz, I., eds., Geological Society of America Memoir 158, p. 161-176.

Glover, L., III, and Gates, A.E., 1987, Alleghanian orogeny in the central and southern Appalachians: Geological Society of America Abstracts with Programs, v. 19, p. 86.

Glover, L., III, Speer, J.A., Russell, G.S., and Farrar, S.S., 1983, Ages of regional metamorphism and ductile deformation in the central and southern Appalachians: Lithos, v. 16, p. 223-245.

Gwinn, V.E., 1970, Kinematic patterns and estimates of lateral shortening, Valley and Ridge and Great Valley provinces, central Appalachians, south-central Pennsylvania, *in* Studies of Appalachian Geology: Central and Southern, Fisher, G.W., Pettijohn, F.J., Reed, J.C., Jr., and Weaver, K.N., eds., Interscience Publishers, New York, p. 127-147.

Hatcher, R.D., Jr., 1987, Tectonics of the southern and central Appalachian internides: Annual Review of Earth and Planetary Sciences, v. 15, p. 337-362.

Hatcher, R.D., Jr., and Zeitz, I., 1980, Tectonic implications of regional aeromagnetic and gravity data from the southern Appalachians, *in* The Caledonides in the U.S.A., volume 2, Wones, D.R., ed., Memoir Virginia Polytechnic Institute and State University, p. 235-244.

Hepburn, J.C., Lux, D.R., and Hill, M., 1987, 40Ar/39Ar Carboniferous cooling ages from the Nashoba terrane, eastern Massachusetts: Geological Society of America Abstracts with Programs, v. 19, p. 18.

Hill, J.D., 1986, Granitoid plutons in the Canso area, Nova Scotia: Current Research, Part A, Geological Survey of Canada, Paper 86-1A, p. 185-192.

Horton, J.W., Jr., Sutter, J.F., Stern, T.W., and Milton, J.D., 1987, Alleghanian deformation, metamorphism, and granite emplacement in the central Piedmont of the southern Appalachians: American Journal of Science, v. 287, p. 636-660.

Kulander, B.R., and Dean, S.L., 1986, Structure and tectonics of central and southern Appalachian Valley and Ridge and Plateau provinces: American Association of Petroleum Geologists Bulletin, v. 70, p. 1674-1684.

Lapham, D.M., and Bassett, W.A., 1964, K-Ar dating of rocks and tectonic events in the Piedmont of southeastern Pennsylvania: Geological Society of America Bulletin, v. 75, p. 661-668.

Ludman, A., 1986, Timing of terrane accretion in eastern and east-central Maine: Geology, v. 14, p. 411-414.

Mitchell, J.P., and Forsythe, R.D., 1988, Late Paleozoic noncoaxial deformation in the Green Pond outlier, New Jersey Highlands: Geological Society of America Bulletin, v. 100, p. 45-59.

Mitra, G., 1979, Ductile deformation zones in Blue Ridge basement rocks and estimation of finite strain: GSAB, Part I, v. 90, p. 935-951.

Mitra, S., 1988, Three-dimensional geometry and kinematic evolution of the Pine Mountain thrust system, southern Appalachians: Geological Society of America Bulletin, v. 100, p. 72-95.

Nance, R.D., 1986, Late Carboniferous tectonostratigraphy in the Avalon terrane of southern New Brunswick: Maritime Sediments and Atlantic Geology, v. 22, p. 308-326.

Nelson, K.D., and others, 1985, New COCORP profiling in the southeastern United States. Part I: Late Paleozoic suture and Mesozoic rift basins: Geology, v. 13, p. 714-718.

Nelson, K.D., McBride, J.H., and Arnow, J.A., 1986, Deep reflection character, gravity gradient, and crustal thickness variation in the Appalachian orogen: relation to Mesozoic extension and igneous activity: Geological Society of America Abstracts with Programs, v. 18, p. 704-705.

Nickelsen, R.P., 1979, Sequence of structural stages of the Alleghany orogeny at the Bear Valley strip mine, Shamokin, Pennsylvania: American Journal of Science, v. 279, p. 225-271.

O'Hara, K., 1986, Tectonic implications of late Paleozoic metamorphism in southeastern New England: Geology, v. 14, p. 430-432.

O'Hara, K., and Gromet, L.P., 1985, Two distinct late Precambrian (Avalonian) terranes in southeastern New England and their Late Paleozoic juxtaposition: American Journal of Science, v. 285, p. 673-709.

Palmer, A.R., 1983, The decade of North American geology 1983 time scale: Geology, v. 11, p. 503-504.

Price, R.A., and Hatcher, R.D., Jr., 1983, Tectonic significance of similarities in the evolution of the Alabama-Pennsylvania Appalachians and Alberta-British Columbia Canadian Cordillera, *in* Contributions of the Tectonics and Geophysics of Mountain Chains, Hatcher, R.D., Jr., Williams, H., and Zeitz, I., eds., Geological Society of America Memoir 158, p. 149-160.

Rast, N., 1984, The Alleghanian orogeny in eastern North America, *in* Variscan Tectonics of the North Atlantic Region, Hutton, D.H.W., and Sanderson, D.J., eds., Geological Society Special Publication 14, p. 197-217.

Rast, N., and Kohles, K.M., 1986, The origin of the Ocoee supergroup: American Journal of Science, v. 286, p. 593-616.

Russell, G.S., Russell, C.W., and Farrar, S.S., 1985, Alleghanian deformation and metamorphism in the eastern North Carolina Piedmont: Geological Society of America Bulletin, v. 96, p. 381-387.

Secor, D.T., Snoke, A.W., Bramlett, K.W., Costello, O.P., and Kimbrell, O.P., 1986b, Character of the Alleghanian orogeny in the southern Appalachians: Part I. Alleghanian deformation in the eastern Piedmont of South Carolina: Geological Society of America Bulletin, v. 97, p. 1319-1328.

Secor, D.T., Snoke, A.W., and Dallmeyer, R.D., 1986a, Character of the Alleghanian orogeny in the southern Appalachians: Part III. Regional tectonic relations: Geological Society of America Bulletin, v. 97, p. 1345-1353.

Simpson, R.W., Shride, A.F., and Bothner, W.A., 1980, Offshore extension of the Clinton-Newbury and Bloody Bluff fault systems of northern Massachusetts, *in* The Caledonides in the U.S.A., volume 2, Wones, D.R., ed., Memoir Virginia Polytechnic Institute and State University, p. 229-233.

Sinha, A.K., and Zeitz, I., 1982, Geophysical and geochemical evidence for a Hercynian magmatic arc, Maryland to Georgia: Geology, v. 10, p. 593-596.

Smith, P.V., and Barosh, P.J., 1983, Detailed structure of a small segment of the Bloody Bluff fault zone in the Concord quadrangle, Massachusetts: Geological Society of America Abstracts with Programs, v. 15, p. 130.

Song, T.S., and Hill, M.L., 1988, The Martic zone, southeastern Pennsylvania: A terrane boundary?: Geological Society of America Abstracts with Programs, v. 20, p. 71.

Steltenpohl, M.G., Hooper, R.J., and Hatcher, R.D., Jr., 1988, Kinematics of fault zones along the east and west flanks of the Pine Mountain terrane, Alabama and Georgia: Geological Society of America Abstracts with Programs, v. 20, p. 317-318.

Thomas, M.D., 1983, Tectonic significance of paired gravity anomalies in the southern and central Appalachians, *in* Contributions to the Tectonics and Geophysics of Mountain Chains, Hatcher, R.D., Jr., Williams, H., and Zeitz, I., eds., Geological Society of America Memoir 158, p. 113-124.

Thomas, W.A., 1977, Evolution of Appalachian-Ouachita salients and recesses from reentrants and promontories in the continental margin: American Journal of Science, v. 277, p. 1233-1278.

Tull, J.F., 1984, Polyphase late Paleozoic deformation in the southeastern foreland and northwestern Piedmont of the Alabama Appalachians: Journal of Structural Geology, v. 6, p. 223-234.

Valentino, D.W., 1988, Rosemont shear zone, central Appalachian Piedmont: dextral displacement of the Willmington Complex: Geological Society of America Abstracts with Programs, v. 20, p. 77.

Vauchez, A., 1987, Brevard fault zone, southern Appalachians: A medium-angle, dextral, Alleghanian shear zone: Geology, v. 15, p. 669-672.

Vauchez, A., Kessler, S.F., Lecorche, J.P., and Villeneuve, M., 1987, Southward extrusion tectonics during the Carboniferous Africa-North America collision: Tectonophysics, v. 142, p. 317-322.

Webb, G.W., 1969, Paleozoic wrench faults in Canadian Appalachians, *in* North Atlantic - Geology and Continental Drift, Kay, M., ed., American Association of Petroleum Geologists Memoir 12, p. 754-786.

Wetherill, G.W., Tilton, G.R., Davis, G.L., Hart, S.R., and Hopson, C.A., 1966, Age measurements in the Maryland Piedmont: Journal of Geophysical Research, v. 71, p. 2139-2155.

Wintsch, R.P., and Sutter, J.F., 1986, A tectonic model for the Late Paleozoic of southeastern New England: Journal of Geology, v. 94, p. 459-472.

Yeo, G.M., and Gao Ruixing, 1986, Late Carboniferous dextral movement on the Cobequid-Hollow fault system, Nova Scotia: evidence and implications: Current Research, Part A, Geological Survey of Canada, Paper 86-1A, p. 399-410.

Zen, E., 1983, Exotic terranes in the New England Appalachians - limits, candidates, and ages: A speculative essay, *in* Contributions to the Tectonics and Geophysics of Mountain Chains, Hatcher, R.D., Jr., Williams, H., and Zeitz, I., eds., Geological Society of America Memoir 158, p. 55-81.

Figure Captions

Figure 1: 1a shows the physiographic provinces of the southern and central Appalachians refered to in text. 1b shows the various tectonic features referred to in the text and Table 1: 1. Long Range-Cabot fault, 2. Catamaran fault, 3. Bellisle fault, 4. Peekaboo-Berry Mills fault, 5. Clover Hill fault, 6. Harvey-Hopewell fault, 7. Hollow fault, 8. Cobequid-Chedabucto fault, 9. St. John fold and thrust belt, 10. Norumbega fault, 11. Kingman fault, 12. Kearsarge-central Maine synclinorium, 13. Bloody Bluff fault, 14. Lake Char / Clinton-Newbury fault zone, 15. Honey Hill fault, 16. Hope Valley shear zone, 17. Narragansett Basin, 18. Martic zone, 19. Rosemont shear zone, 20. Baltimore gneiss domes, 21. Hylas shear zone, 22. Brookneal shear zone, 23. Bowens Creek shear zone, 24. Hollister shear zone, 25. Nutbush Creek shear zone, 26. Modoc zone, 27. Kings Mountain Belt shear zone, 28. Brevard zone, 29. Goats Rock shear zone, 30. Towaliga shear zone, 31. Roanoke Recess, 32. Blue Ridge thrust, 33. Pulaski thrust, 34. Grandfather Mountain Window, 35. Pine Mountain thrust, 36. western limit of Alleghanian deformation. Circled numbers are cross-section lines from Table 1.

Figure 2: Age and relative position of Alleghanian structures in the Appalachian Mountains.

Figure 3: Method of determining the motion with respect to undeformed North America of a point (5) on the outboard side of the orogen that is displaced along an orogen-parallel strike-slip fault.

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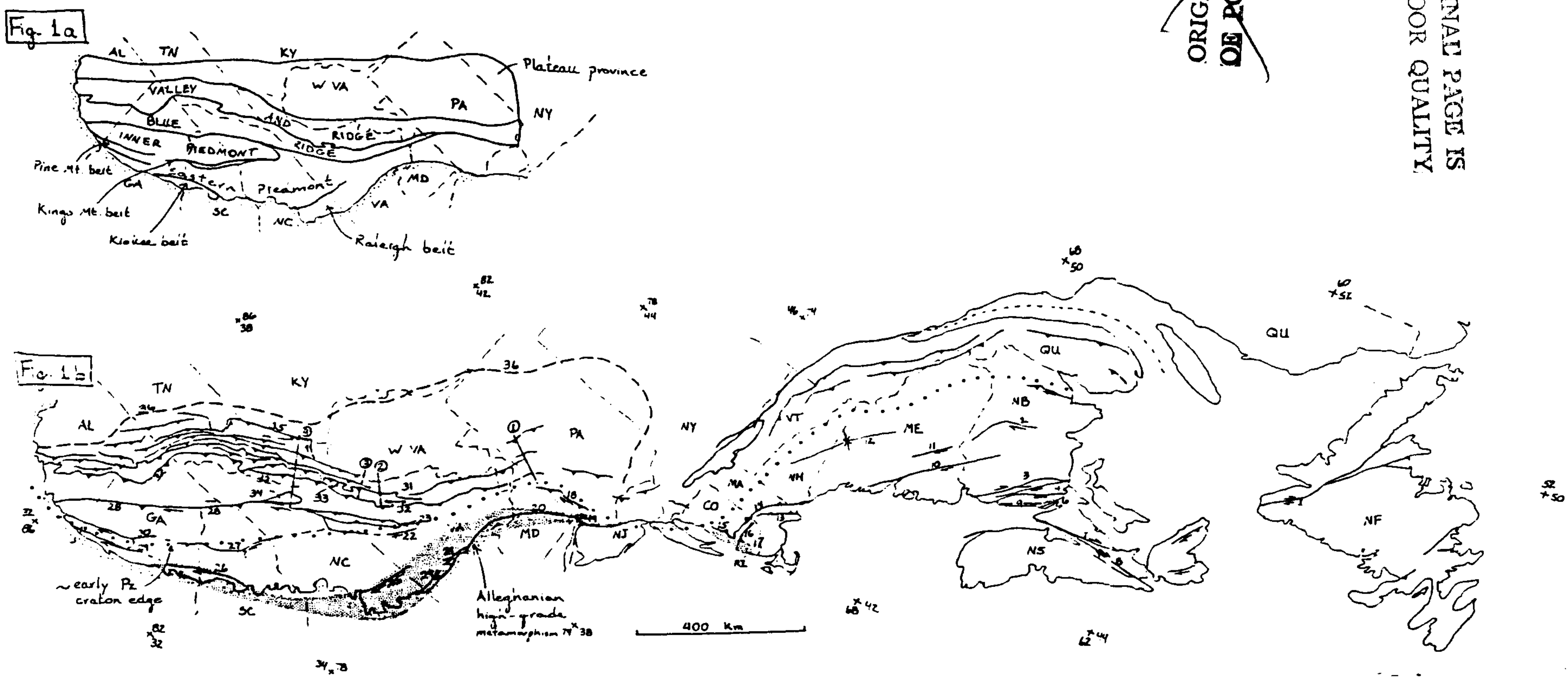

Fig. 1

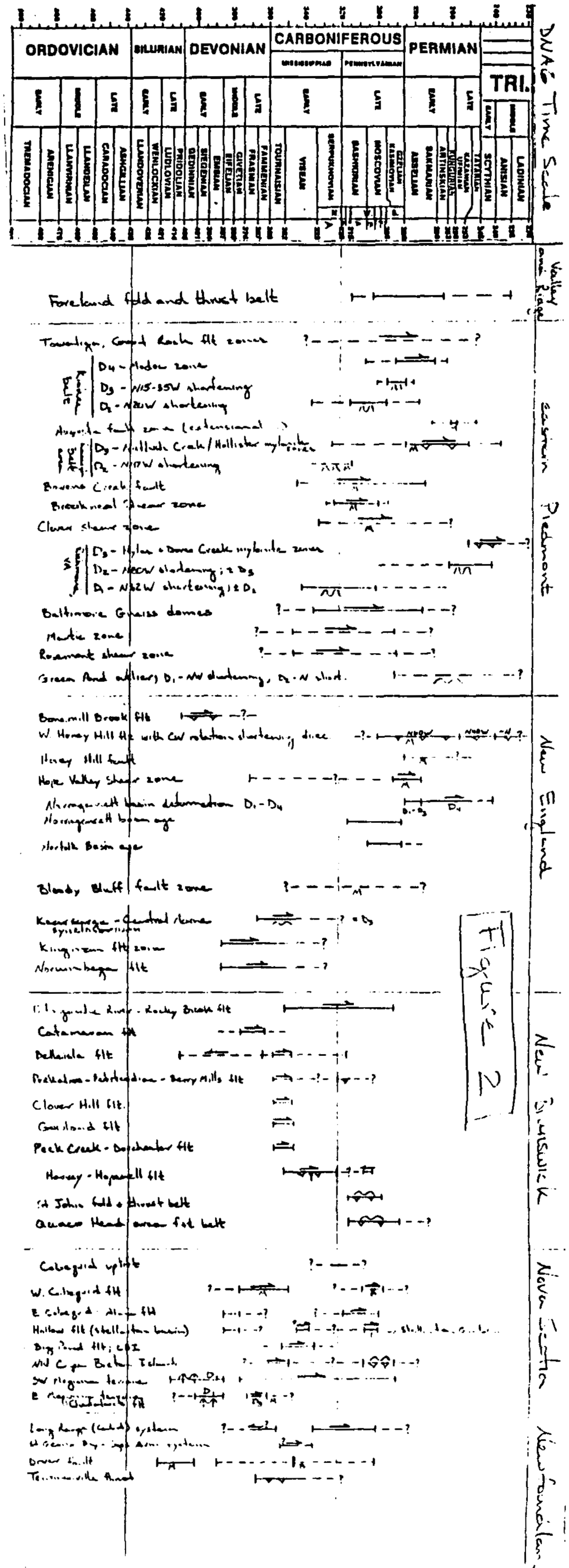

Figure 2

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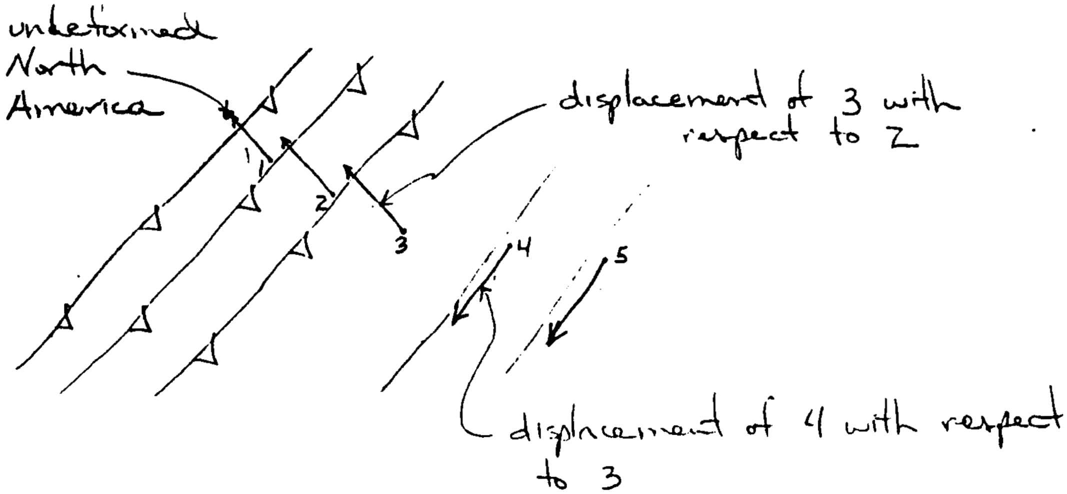
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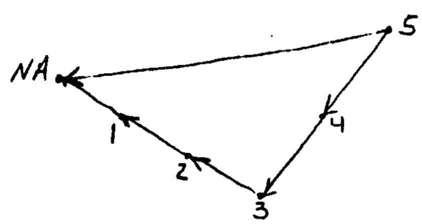
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Figure 3

Table 1: Alleghanian displacement rates

	Name of structure	Displacement direction	Displacement magnitude min./average/max. (km)	Displacement timing beginning/end (Ma)	Disp. time interval min./pick/max. (Ma)	Displacement rate (mm/yr)
orogen-parallel displacements	Catamaran fault [1]	S73W	7 / 12 / 16	374±18 / 360±10 [A]	--/ 14 / 42 [a]	.82 ±.65
	Bellisle fault [2]	S44W	32 / 48 / 64	355 ±9 / 345 ±15 [A]	--/ 10 / 34	4.80 ±4.03
	Peekaboo-Berry Mills fault [2]	S44W	32 / 35 / 38	355 ±9 / 345 ±15 [A]	--/ 10 / 34	3.50 ±2.56
	Clover Hill fault [2]	S57W	9 / 12 / 15 [b]	355 ±9 / 345 ±15 [A]	--/ 10 / 34	1.20 ±.94
	Cobequid-Chedabucto fault [2]	west	157 / 210 / 263 [b]	374 ±18 / 296 ±10 [A,B]	--/ 78 / 106	2.69 ±1.21
	Harvey-Hopewell fault [2]					
	dextral displacement	S42W	60 / 80 / 100 [b]	350 ±10 / 320 ±20 [A]	--/ 30 / 60	2.67 ±1.67
	sinistral disp.	N42E	greater than 16	310 ±15 / 300 ±12 [A]	--/ 10 / 37	>1.60 ±1.17
	Long Range fault [2]	S51W	105 / 181 / 257	333 ±22 / 300 ±12 [A]	--/ 33 / 67	5.48 ±3.92
	Hollow fault [3]	S75W	26 / 35 / 44 [b]	345 ±15 / 298 ±12 [A]	--/ 47 / 74	.74 ±.39
	Hope Valley shear zone [4]	south	25 / 45 / 65	290 / 276 [A,B]	--/ 14 / 21 [c]	3.21 ±2.02
	Rosemont shear zone	S45W	greater than 25 [5]	330 ±16 / 290 [6,C,d]	--/ 40 / 56	>.63 ±.18
	Baltimore gneiss domes	S65W	112 / 150 / 188 [7,b]	310 ±25 / 280 [8,C,d]	--/ 30 / 55	5.00 ±2.96
	Brookneal shear zone [9]	S50W	greater than 17	324 ±3 / 300 ±5 [B,C]	--/ 24 / 32	>.71 ±.18
	Nutbush Creek shear zone	S20W	120 / 160 / 200 [10,b]	313-285 / 251-238 [11,B,C]	34 / 54 / 75	$2.96^{+2.92}_{-1.36}$
	Modoc zone	S67W	18 / 25 / 32 [12,b]	292 ±15 / 268 ±5 [13,B,C]	--/ 24 / 44	1.04 ±.63
orogen-normal displacement	Valley and Ridge					
	section 1 [14]	N67W	103 / 138 / 173 [b,e]	303 ±13 / 265-235 [f]	--/ 38 / 81	3.63 ±2.36
	section 2 [15,16]	N50W	150 / 208 / 267 [e]	303 ±13 / 265-235	--/ 38 / 81	5.47 ±3.55
	section 3 [15,16]	N28W	213 / 259 / 305 [e]	303 ±13 / 265-235	--/ 38 / 81	6.82 ±4.03
	section 4 [15,17,18]	N35W	202 / 269 / 336 [b,e]	303 ±13 / 265-235	--/ 38 / 81	7.08 ±4.59

1. Anderson, 1972
2. Webb, 1969
3. Yeo and Gao Ruixing, 1986
4. O'Hara and Gromet, 1985
5. Valentino, 1988
6. Lapham and Bassett, 1964
7. Glover and Gates, 1987
8. Wetherill and others, 1966
9. Gates and others, 1986
10. Druhan and others, 1988
11. Russell and others, 1985
12. Dennis and Secor, 1985
13. Secor and others, 1986b
14. Gwinn, 1970; Mitra, 1979
15. Kulander and Dean, 1986
16. Bartholomew, 1987
17. Boyer and Elliot, 1982
18. Mitra, 1988

Structures and cross-section lines are located on Fig. 1. Strike-slip displacement directions are dextral unless noted otherwise and give the transport direction with respect to North America of rocks outboard the fault.

A. stratal timing constraints
B. intrusive ages
C. metmorphic ages

a. minimum is given only when the displacement begins within one time interval and ends within another (see text)
b. min. and max. are 25% more and 25% less than a single value given in the cited reference
c. max. is best pick interval plus 50%
d. deformation assumed to occur with last major thermal event
e. magnitudes are the sum of displacements across discret parts of the Valley and Ridge
f. see text for explanation of this timing

Appendix 4

PHANEROZOIC TECTONIC EVOLUTION OF SOUTHERN NORTH AMERICA-CARIBBEAN-NORTHERN SOUTH AMERICA: A PRELIMINARY ANALYSIS

R. Speed
Department of Geological Sciences
Northwestern University
Evanston, IL 60208

April 12, 1988

Introduction

This study takes up the Phanerozoic kinematic evolution of the region between the North American and South American cratons (Fig. 1) with the ultimate goal of understanding displacement rate fields and the position and character of principal displacement zones and their temporal characteristics (steadiness, episodicity, abruptness of kinematic changes). The geographic region includes the North America craton margin from Georgia to Arizona, and to the south, lithospheres of transitional, oceanic, and arc origin, and the continental margin of northern South America. These are considered together because the two cratons were contiguous at least once in Phanerozoic history.

In this report, we outline progress in organizing a tectonic history of the region and data on timing, direction, and magnitude of displacements. The following section gives a time series of major kinematic phenomena. Later sections elaborate on some of these. Future reports will take up all kinematic events and give refinements of displacement fields in each event.

Timetable of Phanerozoic Kinematic Events

650?-520 ma (Late Precambrian—Cambrian): rifting and breakup in and around Proterozoic North America; drifting away of peripheral continental fragments (unidentified); development of a passive margin that extended around much of present North America, including the southern margin from Georgia west to at least eastern Texas (Fig. 2).

520-300 ma (Late Cambrian—Late Mississippian): maintenance of a passive margin in southern North America with heterogeneous subsidence rates; fringe effect of southern most recognized convergent Taconian events (Late Ordovician) in Georgia at eastern end of southern North American margin (Fig. 2).

300-265 ma (Late Mississippian—mid-Permian): right oblique collision of southern North America with Gondwana (South America and Africa) (Fig. 3,4); North America was the downgoing plate; includes emplacement of terranes and North American foreland deformation and subsidence.

265-220 ma (mid-Permian—Late Triassic): mid-continental emergence in western Pangea (Fig. 4); no recognized horizontal motions.

220-140 ma (Late Triassic—Late Jurassic): rifting of central Pangea, drifting of North America and South America, and development of continental microplates and of diverse passive margins; these include the Gulf coastal region, northern South America, and possibly at several positions in between; development of subcontinental subduction zone below western end (Mexico, Arizona).

140-85 ma (Late Jurassic—mid-Cretaceous): subsidence of early Mesozoic passive margins and continued drifting apart of North America and South America, accompanied by spreading in Caribbean basins and/or progressive invasion by Pacific lithosphere; includes widespread midplate magmatism in today's Caribbean basin lithospheres at 110-85 ma with large crustal thickening and probably, extension.

85 ma-present (mid-Cretaceous—present): end thermal contraction/subsidence of passive margins; end North America-South America drifting and begin small incoherent relative motions of NA and SA plates; westward displacement of NA and SA plates relative to the Caribbean plate, resulting in oblique convergent or transgressive overriding of the Caribbean plate relative to NA and SA (Fig. 8); development of Cayman spreading center between 30 and 50 ma (Fig. 6).

Discrete Kinematic Regimes

From the foregoing timetable, four sequential kinematic regimes can be identified:

1. Passive margin development in southern North America in late Precambrian - Cambrian time.
2. North America-South America collision in late Paleozoic time: the southern Appalachian-Ouachitan-Marathonian orogenics and their continuation in Mexico.
3. Breakup of Pangea, divergence of North American-South American plates in Mesozoic time, and the development of intervening tracts.
4. Development of Caribbean plate boundary zones relative to North America and South America in the Cenozoic, especially the Caribbean-South American oblique collision zone.

The organization of this report follows this scheme of four regimes.

Topical Highlights

The fundamental shape of southern continental North America was created by intracontinental divergence at the beginning of Paleozoic time. The edge of North America so formed has been the locus of both divergent and convergent kinematic events in subsequent times.

The structures created in the multiphase tectonic evolution of southern North America have orientations controlled at least partly by preexisting structures. The effect of inherited discontinuities or anisotropy appears more important in structural orientation than farfield displacements.

Although the Mesozoic divergence between North and South American cratons appears to have included uniform and smoothly-varying displacement-rate fields, the intervening region may have undergone complex kinematics, including multiple displacement zones of diverse orientation and rotations.

Each of the major kinematic regimes affecting southern North America appears to have been substantially oblique and included ratios of margin-parallel to margin-normal components ≥ 1.

Collisional regimes in southern North America and northern South America have apparently had partitioned displacements between more nearly continuous margin-normal components and more widely discrete margin-parallel or margin-oblique components. Margin-normal displacement and gradients may have been largely controlled by ramp effects of the passive continental margins.

Each of the major displacement zones occupies belts of great width, at least 200 kms.

Early Passive Margin (Precambrian—Cambrian) of Southern North America

A passive margin developed as a southern boundary of North America near the beginning of Paleozoic time (Fig. 2). This event was of great importance because 1) it was a segment of divergent plate tectonics that essentially blocked out the North American continent as it exists today, and 2) it created two structures: the edge of the sialic continent and the hinge of the subsiding shelf, that are reference lines for subsequent kinematics at the continent's margin. The evidence for such rifting and continental breakup is stratigraphic and structural — the outward transition from platformal to thickened shelfal strata and subjacent rift basins. The position of the cryptic sialic edge, tectonically buried in subsequent convergent regimes, is suggested by isotopic gradients in plutonic rocks and locally, by gravity gradients, magnetic pattern changes, and deep reflection profiling (Rankin, 1975; Thomas, 1977, 1983, 1985; Zietz, 1982; Brown and others, 1981; Lillie and others, 1983).

The model for the edge of Precambrian North America shown in Figure 2 (Thomas, 1983, 1985) implies the following: 1) NW-SE-extension affected a broad zone of eastern and southern North America during the rift phase; 2) a major change in mechanism of breakup occurred at a corner in southeastern North America: a mainly rift margin to the northeast and a mainly transform margin to the southwest, at least as far as Mexico; and 3) apparent diachroneity in breakup; completion early in Cambrian time to the northeast of the corner but slightly later, end middle Cambrian, in the Ouachita region.

The evidence for long transforms in southern North America is the magnetic pattern that suggests a steep ramp at the edge of continental crust and the paucity of Cambrian subsiding shelf strata. Both features imply the absence of a shallow asthenospheric welt leading to greatly stretched continental crust and to a spreading ridge as existed in eastern North America and other mainly rifted margins. Such a correlation between structure and rift vs.transform segments has a good analog in the Mesozoic passive margin of eastern Canada (Keen and others, 1986).

The apparent diachroneity can be explained by the duration of slip on the transform segment required to bring the oceanic lithosphere created in Texas (Fig. 2) past the Ouachita region. If the full spreading rate was 2 cm/yr, the Ouachita region would have been fronted by continental lithosphere for some 20 ma longer than the rifted southeastern Appalachians, even though the onset of breakup was synchronous.

Future work will attempt to relate structures and displacement fields within the craton to this passive margin event.

Collision of North America-South America in Late Paleozoic Time

Gondwana (South America plus Africa) collided with North America and Europe in late Paleozoic time, as indicated by Paleozoic and Mesozoic cratonal polewandering paths and the existence of margin-parallel orogenic belts in southern North America (southern Appalachian, Ouachita, Marathon, Figure 3) and Hercynian belts in southern Europe and northwestern Africa. In southern North America, the collision affected the former passive margin by the emplacement of exotic terranes above the continental edge, the telescoping by imbricate thrusting of outer over inner continental margin rocks, and the subsidence of foreland basins.

The evidence for the postcollisional longitudinal position of Gondwana or the current South American-African boundary (South Atlantic MOR) relative to North America is pliant: 1) the convenient fit of Florida (a probable exotic terrane) in the corner of Gondwana and 2) the distribution of certain faunal provinces in Permian beds, that cover the collision zone (Fig. 4, from Ross, 1979). The crystalline exotic terranes of southern North America outboard of the orogenic belts (Sabine, Wiggins, Suwahnee, Figure 3) are known only from drilling and geophysics. It is a major argument whether the Piedmont terrane, of the southern Appalachian, also crystalline, is exotic or North American basement plus metamorphosed cover. It is here argued that the core allochthons of the exposed Ouachita and Marathon orogenic belts may also be exotic and far-traveled relative to north America.

The extent and position of the collision zone into Mexico are a topic of future investigation in this project.

The essentials of the megastructure of the collision zone, depicted in Figure 3, are discussed below, in landward sequence.

1. The **crystalline terranes** (Sabine, Wiggins, Suwahnee) include one or more late Paleozoic magmatic arcs as indicated by volcanic, plutonic, and metamorphic rocks from drill holes and related magnetic anomalies and by tuff of southerly provenance in late Paleozoic landward marginal basins (Buffler and others, 1988; Thomas, 1985); the magmatic are probably surmounted northern South America and implies that North America was the downgoing plate.

2. The **Ouachita and Marathon terranes** consist of highly scrambled lower Paleozoic strata of deep basinal affinity and very low metamorphic grade. Stratigraphic evidence (Thomas, 1985) indicates these contain some continental debris. It permits the idea they accumulated on an oceanfloor north of northern South America and were accreted to South America upon its transit with a northerly component above a subduction zone. These allochthons may thus represent exposures of the South American forearc (Viele, 1979; Ross, 1985). The forearc, which may extend in the subsurface along the entire orogenic belt, preceded the crystalline terranes in the runup on the North American margin. We suggest the possibility that the Talledega slate belt of the southern Appalachians could be or include the easterly continuation of the Gondwana forearc.

   The alternative to the forearc model is the derivation of strata from basins at the toe of the North American slope, implying parautochthoneity. This idea seems less apt because it requires South America to have been close to North America throughout much of the Paleozoic, countermanding pole-wandering latitudes, unless it is supposed that local, shoaled, continental fragments occurred in the intervening ocean basin.

3. The **foreland thrust belts** take on highly different forms northeast and west of the Alabama corner (Fig. 3). To the northeast in the southern Appalachians the belt is comprised of an imbricate stack of NW-verging slices of thick, coherent lower Paleozoic shelf strata together with piggybacked outer (Carboniferous) foreland basin sediments. To the west,

the foreland belts of the Ouachita and Marathon orogens consist of imbricate stacks of upward-shallowing late Paleozoic turbidites of great thickness (10 km or so). These have varied NE and NW vergences (Fig. 3). The Ouachita and Marathon imbricate stacks apparently include none of and entirely overrode the lower Paleozoic shelf cover which is thought to have been thin and of minor extent in this area (Thomas, 1985).

The spatial change in foreland thrust belts in the collision zone may be explained by the following alternative hypotheses:

a) in the Ouachita-Marathon region, the foreland thrust belt turbidites are really outer foreland basin strata and the exotic terranes have overridden the real foreland thrust belt, if one existed; in this case, the edge of North America may lie well south of its position shown in Figure 2 in the Ouachita belt.

b) the Ouachita foreland thrust belt turbidites were deposited in a remnant continent-ocean gap before a diachronous east-to-west collision imbricated and transported them above the continental margin (Thomas, 1985); in this case, the difference is partly explained by the Ouachitan collision at a transform segment of the former passive margin where thick shelf strata did not exist.

4. The **foreland basin** is an asymmetric flexure of underlying continental lithosphere, almost certainly caused by tectonic vertical loading at positions seaward of the basin's outer edge by the allochthonous terranes and foreland thrust belts. As the load migrated landward, the foreland basin flexure did likewise. The outer (seaward) basin fill was imbricated with underlying shelf strata and included in the foreland thrust belt in the southern Appalachian orogen. In the Ouachita-Marathon orogens, outer foreland basin strata were apparently detached from the shelf strata and imbricated alone as the foreland thrust belt. Landward of the deformation front, the foreland basin fill was undeformed though the collision zone, except for down-to-the-basin (seaward) normal faulting.

The foreland basin is of great importance in kinematic analysis because it indicates the timing of loading (encroachment and sticking of allochthons), gives hints as to thickness and shape of the allochthons by the amplitude and width of basin flexure, and records the reactivation of earlier structures in the takeup of displacement during flexure.

The overall late Paleozoic foreland basin is nonuniform in duration of activity, as indicated by ages of onset and completion of subsidence on Figure 3. The implication is an east-to-west migration of the first and last effects of loading, and recorded in parautochthonous sediments of seaward provenance. The lateral migration duration was about 65-70 ma.

The foreland basin is highly heterogeneous in thickness of fill on strike in the collision zone (Fig. 3), from a few km to >10 km. This might be explained either by large variations on strike of the thickness of the load and/or contemporary breakup or reactivation of older structures in the loaded lithosphere. The former might be thought of as a series of huge ridges ramping up normal to the continental edge. The latter may reflect the development of fault-bounded basins (departing from the flexure of an elastic plate) with enhanced subsidence due to preferential sediment trapping. The latter explanation seems to be demonstrated in the Arkoma Basin and several basins of the Marathon region (Thomas, 1985; Ross, 1985). At both places, at least some basin-bounding faults are thought by these authors to be inherited from the Precambrian-Cambrian breakup.

The displacement field of the late Paleozoic collision belt is nonuniform. The most evident structural features reflecting displacement directions are the horizontal component of the principal contraction (minimum eigenvector) and the vergence of the foreland thrust belts. From an initial look, the displacement gradients are everywhere normal to the continental margin (to the craton-shelf hinge) and the displacements directed landward. The range of directions is at least 45°. The local homogeneity of contraction trends normal to the margin suggests such strains are not rotational about vertical axes. It is unlikely that such directions each represent the local translation vector of segments of the leading edge of South America. It is more probable that the foreland contraction is imposed by the dip of the continental ramp the strata are moving against and up. The contraction would thus reflect the zigzag edge of southern North America and indicate only that South America's translation had a N to NW component. There is evident noncompatibility in the displacement gradients recorded by the foreland thrust belts, and this must be taken up by strike-slip and/or normal faulting between adjacent thrust belt panels. Such structures will be searched for in further work on this project.

To summarize evidence for the translation direction(s) of South America relative to North America, we have 1) the margin-normal foreland contraction and polar-wandering which indicate a northerly component and 2) the diachroneity of foreland deformation which indicates a westerly component. Other variables in the reconstruction we shall ultimately attempt are the irregularity in the edge of and the possible nonrigidity of the overriding plate.

Divergence of North American and South America in Mid-Mesozoic Time

Farfield kinematics: The trajectory and rate of the mid-Mesozoic translation between North American and South American cratons are well resolved for the post-breakup duration (Fig. 5). The translation is established from passive margin to passive margin by rotation poles relative to Africa determined by Atlantic isochrons and fracture zones (Sclater and others, 1977; Pindell and Dewey, 1982). The trajectories show migration of South America generally SE and away from North America from breakup in Late Jurassic (165 ma) time to Late Cretaceous (80 ma) time. The displacement magnitude in this interval is about 3200 km; the average rate is about 3.8 cm/yr. The takeup of rotation by South America or(and) Africa relative to North America is uncertain during opening of the South Atlantic (Rabinowitz and LeBreque, 1979). The cratonal divergence episode is followed in Late Cretaceous and Cenozoic time (after 80ma) by small and unsustained relative displacements between the cratons (Fig. 5).

In contrast to the mid-Mesozoic cratonal divergence, the kinematics and tectonic phenomena in the intervening Gulf-Caribbean region (Figs. 1, 6, 7) during the same interval are poorly understood. It is of great interest in this study to determine the evolution of structures in a region where the farfield kinematics are known apriori. The tectonic components, in theory, are lithospheric stretching, volume addition to continental lithosphere from the mantle (diking), seafloor spreading, convergence (= excess divergence) and tangential invasion of Pacific lithosphere discussed in the next section.

Southern North America and Gulf of Mexico

The rifting that began in Pangea during Late Triassic well into Middle Jurassic time (about 220 to 160 ma) created graben in the upper continental crusts of sutured North and South America. The graben were filled by subaerial sediments and basalt and then by salt when adjacent oceans leaked in, mainly in the Middle Jurassic. The crust was also diked by basalt at sites away from the graben (Fig. 6). The spreading that created the Gulf of Mexico basin at the end of the Middle Jurassic (Salvador, 1987) occurred in a closed basin surrounded by rifted continental crust (Fig. 1). The Gulf Basin must therefore be enveloped by a system of rift and transform margins. However, magnetics and internal fractures are too poorly resolved to indicate a spreading direction. The Gulf Basin spreading stranded in North America terranes (Sabine, Wiggins, Suwahnee, Fig. 6) that originally belonged to South America.

The extensive Gulf Coast salt basins are probably not useful in analysis of horizontal kinematics, because the salt lapped over and obscures the orientation of early graben. The salt basins are indeed important, however, in indicating positions of progressive subsidence.

A problem immediately arises in relating graben and dike orientations and the Gulf Basin boundary faults, so far as these are known (Fig. 6), to the mid-Mesozoic farfield translation.

The graben strike parallel to the ancient continental margin, curving around the Gulf embayment. The dikes of the southern Appalachians not in graben strike NW-SE, such that conventional wisdom would infer NE-SW extension. The most precipitous slopes of the Gulf Basin, properly assumed by Salvador (1987) to be transform margins, strike NNW-SSE. These structures do not form a system of rectilinear rifts and transforms under uniform extension, much less have an evident relation to the farfield NW-SE divergence of the cratons.

We conceive of three alternatives to explain such disparities. First, rift basin orientation is controlled by precursor structure, not by regional stress orientation. This means that only graben with NE-SW strike have dipslip motion and all others have oblique-slip with obliquity proportionality to difference from NE-SW strike. It also means that dikes (Fig. 6) of the southern Appalachians not in graben preferentially intruded transform faults (i.e., faults parallel to the translation direction), and that Gulf Basin spreading had significant obliquity to the farfield translation direction.

A second explanation is that the Triassic and Early Jurassic rift phase kinematics of western Pangea differed directionally (that is, different eigenvectors of deformation) from those that controlled the Late Jurassic and Cretaceous Atlantic spreading and cratonal divergence. Therefore, the pre-Late Jurassic motions may have had widely changing directions with time. Further, it is possible that spreading of the Gulf Basin preceded opening of the Atlantic.

The third explanation is that the region between the separating cratons was kinematically discontinuous with the Atlantic during the phase of cratonal translation and/or later phases. This implies either the existence of plate boundaries surrounding the intervening region during the mid-Mesozoic or later, permitting differently oriented eigenvectors during breakup and spreading and/or later rotation of terranes of the region between the craton relative to the Atlantic.

Current appreciation supports the application of all three hypotheses, as follows: 1) the spacewise correlation of strike of Mesozoic graben and older structures in the Gulf Coast and Appalachians (Fig. 6) is unquestionable, implying structural inheritance; in fact, there are a few specific faults where graben walls are said to have been Paleozoic thrust ramps; 2) the idea of different or nonuniform displacement fields before Atlantic spreading is the only way to explain diking in the direction of maximum shear; and 3) there are suggestions that rotation about the vertical has in fact occurred in the Gulf of Mexico region.

Concerning rotations, paleomagnetic data in Yucatan (Fig. 1) indicate $22 \pm 8^\circ$ clockwise since a time in the Permian (Gose and Sanchez-Barreda, 1981). Yucatan is terrane of continental lithosphere that probably belonged to South American Pangea and has been stranded during the South American Mesozoic getaway by spreading in the Gulf of Mexico and Yucatan Basins (Fig. 1). Questions on the paleomagnetic result are the sample sufficiency and the age of rotation, which might be related to the late Paleozoic collision rather than Mesozoic divergence. Further, restorations of the oceanic Gulf of Mexico basin all come up with nonuniform spreading rates on strike, implying rotation in transitional lithosphere on one or both flanks of the basin (Hall and others, 1982; Pindell, 1985; Salvador, 1987). The problem is that the constraints on restoration are so vague that both clockwise and counterclockwise rotations are variously predicted by these studies.

The central Yucatan terrane differs from the prevailing transitional lithosphere and from the northern and eastern terrane margins by significantly greater crustal thickness, as inferred, however, only from gravity data (Buffler and others, 1983). The implication is that either rifting and stretching in mid-Mesozoic Pangea was heterogeneous or that the Yucatan terrane is not part of the Pangea breakup and had an origin at some more distant site, presumably to the NW or SW.

The Yucatan Basin is known only to contain deep oceanic crust, suggestive of an Eocene or older age. Its spreading direction is unknown, as is the character of its boundaries — tectonic or otherwise. It may be a tectonic terrane, an in-situ late (Eocene) oceanic basin, or a relic of Mesozoic crust created in the wake of South America's migration.

Gulf of Mexico Basin to North American-Caribbean Plate Boundary

This poorly understood region is thought to be composed mainly of transitional (stretched and continental) lithosphere from southern Mexico and the Atlantic margin of the Bahama platform, including Cuba (Figs. 1, 7). The exceptions are the continental Yucatan terrane, the oceanic Yucatan Basin, and a contractile collision zone in Cuba and adjacent marine areas. The transitional lithosphere presumably underwent mid-Mesozoic extension during the separation of the American cratons, but dating in marine regions Schlager and others, 1984) shows only that cessation of rifting was in Early Cretaceous or earlier time. Therefore, it is possible that rifting occurred in this region before, during and/or after the late Middle Jurassic formation of the Gulf of Mexico.

A kinematic study of seismic structures in this region has not yet been made, except to say that long NW-SE bathymetric scarps exist. These are stratigraphic disjunctions which may be faults or carbonate platform margins. If faults, their length of straightness and strike suggest transform faults parallel to the mid-Mesozoic cratonal translation (Fig. 5).

At least the northern half of Cuba is presumably underlain by the transitional lithosphere just discussed (Fig. 1) which is the autochthon of the early Cenozoic collision zone with the Greater Antilles arc. Jurassic strata showing progressive subsidence are said to occur in the autochthon (Pardo, 1975) but the dating is too imprecise, as published, to be useful in tracking rifting across the Gulf-Caribbean region.

The orogenic belt of Cuba has had little modern study for political reasons. Current wisdom suggests it is near the western end of the Greater Antilles arc-continent collision zone, now cut by the North American-Caribbean plate boundary of the Cayman Trench (Figs. 1, 7). In Cuba, the arc was active in Late Cretaceous time and is thought to have ridden over a south-facing edge of the transitional lithosphere from Cuba east to the Bahamas. The collision was possibly completed in Cuba by early Eocene time. The contraction in the Cuban orogenic zone is guessed to be NS. An E-W trending early Tertiary (and older?) sedimentary basin exists at the northern Cuban shoreline, thinning northward. It can be interpreted as a foreland basin or transitional lithosphere due to loading by the overriding arc to the south (Angstadt and others, 1985).

Caribbean-Northern South America

Introduction: This section considers the southern boundary of the North American plate and regions southward to northern South America (Fig. 7). The objective is the organization of a framework to analyze displacement fields of the Caribbean relative to the American plates, the collision of Caribbean lithosphere and continental South America, and the mid-Mesozoic divergence of the cratons in the Caribbean basins (the Colombian and Venezuelan Basins, Fig. 7) and within northern South America.

Northern South America: The structure of northern South America relates chiefly to three tectonic regimes: mid-Mesozoic rifting and passive margin development, Cenozoic collision with Caribbean lithosphere, and the late Neogene translation of the Pacificward quarter relatively northeast on the Bocono fault (Fig. 7). We take up here the passive margin development which relates to the Late Jurassic-Late Cretaceous translation trajectory of relative cratonal motions (Fig. 5).

The northern edge of sialic South America as far east as Trinidad is buried tectonically below south-vergent thrust slices in the Cenozoic Caribbean-South American collision zone (Fig.

7). Evidence that the northern edge was a passive margin with approximately EW trend is the existence, sparingly occurring or recognized, of rift-basins along the northern edge (Fig. 7) and the northerly gradient in thickness of Cretaceous shelf deposits, suggesting thermally controlled subsidence. East of Trinidad, the sialic edge joins Atlantic oceanic lithosphere at a depositionally-buried passive margin with NW-SE trend (Fig. 7). This easterly leg of the sialic edge may be a predominantly transform margin, as interpreted by its parallelism with the mid-Mesozoic cratonal translation direction. Similarly, the less clearly identified EW leg west of Trinidad may have been a zigzag series of rifts and transforms.

It is a question whether the northern South American continent began rifting at the same time as crust of the U.S. Gulf Coast or not. Known older Mesozoic deposits are the following (Fig. 7): 1) Middle Jurassic redbeds with pillow basalt in western Venezuela at Siquisiqui (implying rift phase); 2) latest Jurassic carbonates in thrust slices in the collision zone (unit 2, Fig. 8) of eastern Venezuela and Trinidad (implying drift phase somewhere to the north); 3) evaporites of possible Early Cretaceous age in Trinidad (implying rift phase); 4) possible dipping rift basin fill surmounted unconformably by shelf strata seen in seismic sections just south of Trinidad; and 5) drilling and dredging evidence of Jurassic red sediment off the Guyanese coast and Demarara Rise (implying Jurassic rifting). No Triassic strata are known. The drift phase shelf deposition on South America was underway by a time early in the Cretaceous. The upshot of these data is that rifting in South America may have been started and continued after that of the Gulf Coast. Therefore, it follows that rifting may have propagated or jumped southward in Pangea from an initial locus in the U.S. Gulf Coast. This locus was approximately coincident with the late Paleozoic collision zone of the southern Appalachian-Ouachita-Marathon orogens and invites speculation that the earliest extensional structures were localized by reactivation of older structures.

Caribbean-American Plate Motions

A major problem in the analysis of ancient displacements in orogenic zones between the American cratons is the imprecision in the current plate tectonics of the Caribbean region — the positions of major displacement zones and the relative velocities across them.

The Cayman Trough (Fig. 7) contains the only recognized divergent boundary around the Caribbean, but it is almost totally strike slip, including perhaps the greatest ratio of transform to spreading in the world. A problem with the Cayman Trough is the uncertainty in the total length of opening (estimates are between 270 and 1400 km; Macdonald and Holcombe, 1978; Sykes and others, 1982; Rosencrantz and others, 1987) and the duration of spreading (estimates of 10 to 50 ma).

If the length is >1000 km, most of the Caribbean is Farallon plate that has translated east with respect to the Americas from the Pacific; in this case, the Caribbean basin contains no kinematic record of the diverging American cratons because intervening terranes would have been subducted below the Farallon plate. On the other hand, if the length of Cayman spreading is small, the bulk of the Caribbean basin has been created between the American cratons.

Aside from the Cayman Trough, Cenozoic orogenic belts on the Caribbean rim imply a component of convergence all around the Caribbean plate by the existence of accretionary wedges and fold belts that are more extensive and univergent than flower structures. If correct, this means that either the Caribbean "plate" has a nonuniform displacement field (i.e., a nonrigid plate), or the American plates are convergent relative to one another more than NA-Af-SA rotations suggest (Fig. 5), or both.

The Caribbean-South American plate boundary zone is especially poorly defined. Seismicity gives little evidence for the positions and motions of displacement zones (Molnar and Sykes, 1969; Kafka and Wiedner, 1979; Perez and Aggarwal, 1981; Wadge and Shepherd, 1984). Further, the geologic evidence indicates right-oblique convergence and an absence of individual faults with significant displacement (>100 km) in the ocean-continent transitional zone

(Schubert, 1981; Speed, 1985), although Robertson and Burke (1986) try to make a case for very large strike-slip. It seems possible to accommodate no more than an integral 400 km right slip in this zone since the Eocene. Therefore, if the Cayman Trough (Ca-NAm) has had small total spreading (i.e., 400 km), the Caribbean plate may exist as a rigid edifice. If, however, the Cayman displacement is very large, displacement zones must exist within the Caribbean to accommodate the different Ca-NAm and Ca-SAm total offsets in the last 50 ma. If the differential were linearly distributed across the Caribbean, the simple shear strain would be about 1, and an average rigid rotation would be about 20° clockwise (this is what crustal blocks coupled to a ductile substrate would show paleomagnetically).

A more interpretable active regional displacement zone is the Bocono fault (Figs. 7, 8) which is seismically active and has released the northern Andes from South America to move relatively northeast and override the southern Caribbean (Percy and Aggarwal, 1981; Pindell, 1985). The total displacement, measured by offset of structures, is 100-150 km (Schubert, 1981). The duration of Bocono activity is probably <10 ma, although 50 ma is proposed by Stephan (1977). The length of subducted Caribbean lithosphere is insufficient to have caused arc volcanism.

Caribbean Oceanic Plateau

Ninety percent or more of the Caribbean basins west of the Aves Ridge (Fig. 7) are underlain by anomalously thick and shallow crust (≥ 5 km thicker than normal oceanic) that is at least partly composed of basaltic lava, intrusions, lesser evolved lava, and derivative sediments of Late Cretaceous age (Edgar and others, 1973; Donnelly and others, 1973). This oceanic plateau may extend onland in obducted slices, in Haiti (Sen and others, 1988) and to the south in the Netherland Antilles (Beets and others, 1984; Klaver, 1987), the Villa de Cura and Paraguana allochthons, and east to Tobago (Bellizia, 1985; Speed, 1986). The ages of the lava sequences in obducted slices extend the range of volcanism from about 110-80 ma.

The importance of the oceanic plateau is its 1) evidence for convergence at its margins; 2) mid-late Cretaceous kinematics of its development and bearing on North America-South America separation; and 3) a possible record of pre-mid-Cretaceous tectonics below the plateau lava sequence.

The lava sequence indicates there was large volume addition to the Caribbean crust in a wide region over a relatively short time. Lava compositions indicate large volume melting, storage and evolution in high-level magma chambers, and a remarkable range of source analogs: normal MORB to enriched MORB to hotspot types, all without evident systematic spatial distribution.

A patch of normal oceanic crust occurs in the southeastern Venezuelan Basin (Diebold and others, 1981). It is unclear whether this patch is faulted against the prevailing thick crust or whether the oceanic plateau lavas onlap the edge of the normal crust.

The obduction of Caribbean lithosphere above South America caused a major orogenic belt to evolve above the continent's shelf; this is addressed later.

An understanding of what is below the oceanic plateau lavas is fragmentary. Reflection data in the Caribbean basins locally resolve dipping reflectors below it (Ladd and Watkins, 1979; Diebold and others, 1981), and refraction velocities indicate upper oceanic crustal velocities. Below the Cretaceous lavas in the Villa de Cura allochthon (Fig. 8), there are extensive ophiolites (one is early Cretaceous or older; Beck, 1985), a sequence of arc-related mafic rocks that are deformed and upsidedown, and sialic metamorphic fragments of Precambrian or Paleozoic protolith age. The Villa de Cura allochthon is thought to be derived from regions seaward of South America because the subjacent tectonic unit (unit 3, Fig. 8) contains strata accreted from the continental slope and fringing oceanfloor. Together, these features suggest that before the mid-Cretaceous magmatic event, the Caribbean basin contained rift basins with rotated fill, fragments of Pangea, local basins with oceanic crust, and at least one subduction zone and collapsed island arc.

The mid-Cretaceous plateau-forming volcanism can be postulated to record an acceleration in the divergence of the American cratons where the extension was taken up by volume addition from the mantle in distributed conduits rather than at a single spreading center. An analogy is the Columbia Plateau of the Pacific Northwest or the Basin-Range; the difference with the Basin-Range is the latter's development below mainly thick continental crust. The diversity of magma compositions in the Caribbean oceanic plateau can be attributed to contamination by or sourcing in the highly varied lithospheric fragments that made up the pre-mid-Cretaceous Caribbean basin.

There are no constraints with which to calculate the extension through volume addition to the precursor Caribbean lithosphere. The unknowns include the proportions of extrusion to intrusion and of dikes (extension-producing) to sills. It is conceivable that the vague NE-SW magnetic fabric of Caribbean basins (Donnelly, 1973; Ghosh and others, 1984) may reflect extensional mid-Cretaceous dike systems.

A leading alternative to the origin of the Caribbean oceanic plateau proposed above is control by a global or equatorial mid-Cretaceous volcanic event, best known in the western Pacific (Larson and Schlanger, 1976). Although a mechanism has not been put forth for this global event, its occurrence in the Caribbean implies a cause more profound than local lithospheric thinning due to continental divergence.

Caribbean-South American Oblique Collision

To address the structure of northern South America as it may relate to the Mesozoic breakup and pull away from southern North America, the subsequent convergent margin tectonics of northern South America must first be analyzed. Moreover, this Cretaceous-Cenozoic episode includes important kinematic phenomena in its own right. Such an analysis requires a generalized tectonic framework from the geologic and geophysical literature and our own observations, and thus, is model-dependent and almost certainly, to be iterated. Our main sources of information are Bellizia, 1972, 1984; Campos, 1981; Maresch, 1972, 1974; Gonzales de Juana and others, 1980; Beets and others, 1984; Klaver, 1987; Skerlec and Hargraves, 1981; Stephan and others, 1980; C. Beck, 1985; M. Beck, 1986; Silver and others, 1985; Stearns and others, 1982; and Speed, 1985, 1986, and unpublished field work in eastern Venezuela and Trinidad.

Figure 8 shows a new tectonic model of northern South America, packaged in but four tectonic units. The four units are parallel to and above the cryptic EW trending rifted Jurassic passive margin of northern South America. The tectonic units stack upward toward the north, and in a complex way, dip north. The numbers are the age limits within which the associated major displacement zone ceased movement and the fault walls attached. The numbers show a general west to east migration of attachment, and thus, indicate the oblique overriding of the continent by northerly terranes initially at an angle to the continent's EW-trending edge. The region northwest of the Bocono fault is translating northeast and overriding the Caribbean lithosphere (Fig. 8).

The southernmost and lowest unit is Proterozoic South America and its little deformed cover. This unit has undergone displacement only vertically by flexure in the development of a foreland basin at the toe of the orogenic belt (Fig. 8). The amplitude of flexural depression is known to be as great as 10 km in the Cenozoic.

The second lowest unit, the foreland thrust belt, is constituted by imbricated Mesozoic shelf strata, probably deposited on subsided Jurassic-Cretaceous passive margin, and Cenozoic strata of which the older are shelf strata and the younger, piggy-backed foreland basin deposits. The age boundary between older and younger changes with position: late Paleocene in the west and Miocene in the east. This diachroneity of transition from shelf to foreland basin supports the idea of migration west to east of the load on the northern edge of the continent. The load was supplied by the foreland thrust belt but chiefly, by the northern two allochthons.

The displacement field of the foreland thrust belt is under study. In western Venezuela, the narrow thrust belt (Fig. 8) has structures indicating NS contraction of uncertain value. There, much of the thrust belt has probably been overridden by the higher allochthons during Cenozoic out-of-sequence thrusting. In eastern Venezuela, the contraction direction is NNW-SSE and is roughly 0.5 in magnitude (imbrication doubled the thickness of strata). Associated structures (a single set of NW-SE major strike slip cross faults, Riedel and antiriedel shears, and normal faults) give a good fit to horizontal eigenvectors of deformation shown in Figure 9. The right-slip cross faults are probably parallel to the displacement direction of thrust slices (NW to SE) across the belt in eastern Venezuela. This is because they formed during imbrication of thrust slices and separate en-echelon adjacent panels of imbricates; the cross faults, however, seem not to have undergone rotation with deformation as they would if they were not principal displacement surfaces.

The third unit from the bottom (Fig. 8) contains metamorphosed, well-deformed sedimentary rocks that are Mesozoic $\pm$ Cenozoic age and slices of more ancient rock. These are here interpreted to be packaged in an extensive accretionary forearc that propagated ahead and below the highest tectonic unit (Caribbean lithosphere). The contents of unit 3 were accreted to the south and(or) southeast-translating forearc from oceanfloor that fronted the mid-Mesozoic South American passive margin and from the slope and shelf of that margin as well. The metamorphism was caused by deep burial and(or) the overriding of the forearc by the Caribbean lithosphere. The occurrence of pre-Mesozoic sialic masses in unit 3 (example: Early Cambrian Dragon gneiss, Fig. 1) suggests either that the edge of tectonically buried continental crust extends north of unit three's exposure belt or that continental fragments existed north of the passive margin. The deformation of unit 3, where studied in Trinidad and Margarita, shows ductile NS contractions plus rotations about an EW horizontal axis. The total length of contractile displacement within unit 3 is probably very large because horizontal contractions ($\Delta l/lo$) appear to be ≥ 1.0 at the few sites examined.

The highest unit (4) is thought to consist mainly of Caribbean lithosphere that has been thrust with a southeasterly component above an unknown width of the Mesozoic passive margin of South America (Fig. 8). Subjacent units 3 and 2 contain rock progressively accreted to the front and bottom of the Caribbean lithosphere and shipped with it southeasterly above South America.

The basis for the assignment of unit 4 as Caribbean lithosphere is its content, where studied, of thick (up to 5 km) mid- or mid-late Cretaceous basaltic lava plus pelagic and volcanogenic sediment and locally differentiated magmatic rocks. Such rocks are underlain by subunits of diverse origin (ophiolite, continental fragment, island arc), hence terranes, that appear to have been linked together by the mid-Cretaceous volcanic event. Further, it can be suggested that the mid-Cretaceous volcanic event is the same as the one that formed the Caribbean oceanic plateau that occupies most of the deep Caribbean basins north of the continental borderland (north of Figure 8), as discussed earlier.

The importance of the correlation of Caribbean lithosphere in the collision belt and Caribbean basins lies in the following: 1) possible evidence for length of convergent transport across the continental margin (the position of the buried continental edge still has to be determined by some remote means); 2) onland recognition of mechanisms by which the Oceanic Plateau formed and its kinematics; 3) a glimpse at the structure of Caribbean basins before the plateau-forming event; and 4) the possibility of constraining whether present Caribbean lithosphere is mainly indigenous or far-traveled.

Regarding the first item, the sediments of unit 3 are largely of continental provenance but do not permit differentiation among oceanic and continental sites of deposition. They do permit the inference, however, that the Caribbean oceanic plateau developed seaward of their sites and that the southerly transport of Caribbean lithosphere is at least 200 km (100 km outcrop width and assumed 100% contraction of unit 3). This is supported by the quasicontinental initial Sr values in post deformational plutons in unit 3 near Caracas (Fig. 8).

Items 2 and 3 are discussed in a prior section. Concerning item 4, the correlation suggests that Caribbean lithosphere off Venezuela and Colombia is not far-traveled (not from distant sites in the Pacific). This is because units 4 and at least part of 3 were linked by a time late in the Cretaceous (C. Beck, 1985) and because unit three's sediments are partly continental. Therefore, units 4 and 3 were together before any recognized onset of Cayman Trough opening and generally eastward translation of the Caribbean plate relative to the American plates.

The data concerning the displacement-rate field of the Caribbean-South American collision also include sparse paleomagnetic sets that survive statistical tests of precision (Beck, 1986) and for most of these, structural control is poor. The suggestion from paleomagnetism is that a large clockwise rotation (45 to 90°) relative to South America exists in at least some mid-Cretaceous igneous rocks of the Caribbean lithosphere (unit 4, Fig. 1); the rotation occurred before the Cenozic.

To summarize, the kinematic data and suggestions amassed to date are: 1) large southerly closure of the Caribbean lithosphere relative to South America in late Cretaceous and Cenozoic time; at least off western Venezuela, the total closure probably exceeds 200 km; 2) eastward migration of the onset and cessation of tectonic effects in the foreland thrust belt and basin; 3) interpretation of a NW to SE tectonic transport in the eastern foreland belt mainly in Neogene time with a local easterly migration rate of 1 to 1.5 cm/yr; and 4) probable large clockwise rotation of at least parts of the highest unit, the Caribbean lithosphere.

The initial configuration of the Caribbean lithosphere is uncertain relative to an EW-trending edge of mid-Mesozoic South America. Figure 10 shows two alternatives. The convergent boundary may have been at a small angle to the continental edge (Speed, 1985), an hypothesis that requires the paleomagnetic rotations to be products of local ramping. The alternative is that the convergent margin had a NS strike and that Caribbean lithosphere wrapped around at right angles where it overrode the South American margin (Skerlec and Hargraves, 1981).

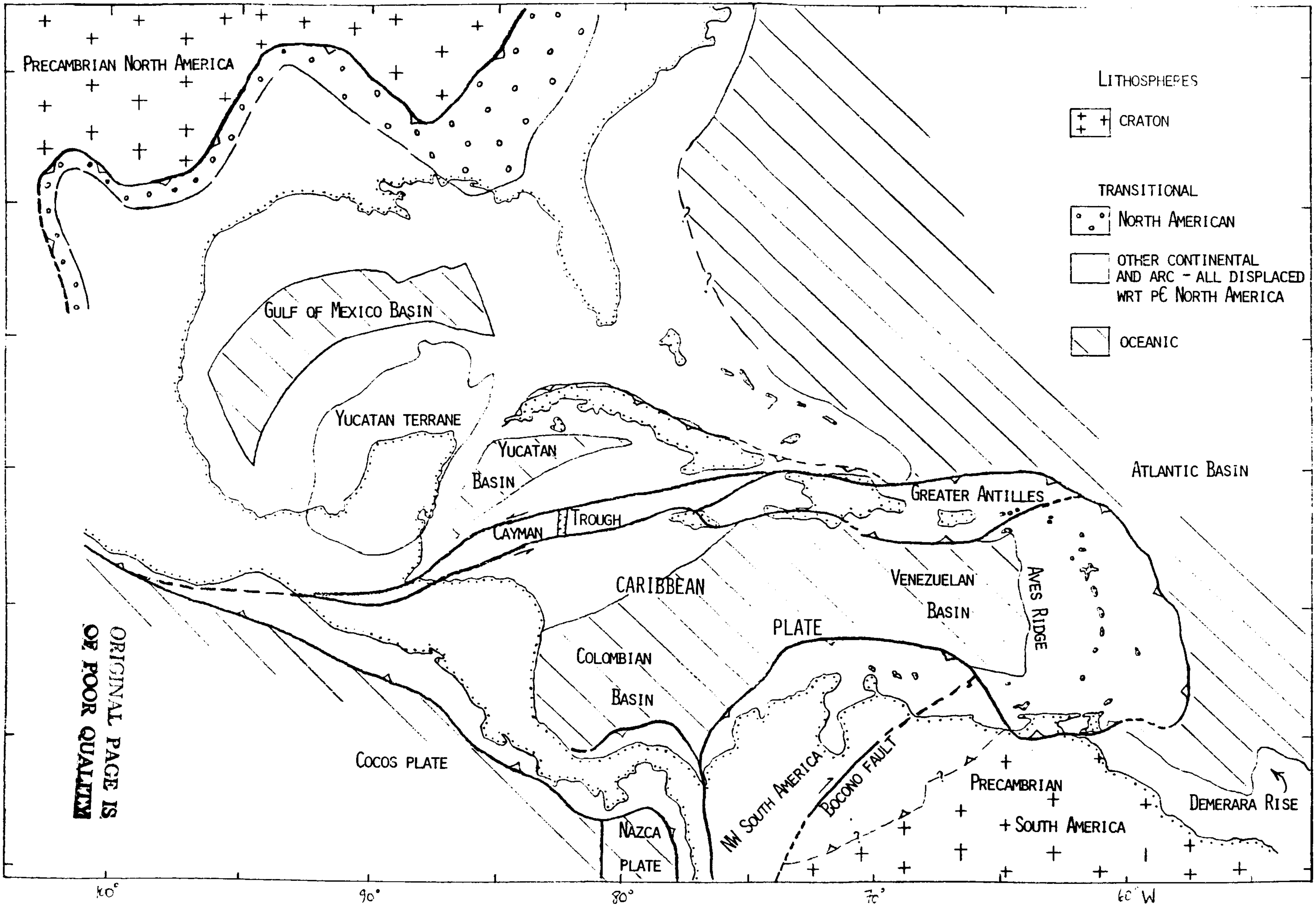

Figure 1: Tectonic map of region from southern North America to northern South America, divided by type of lithosphere.

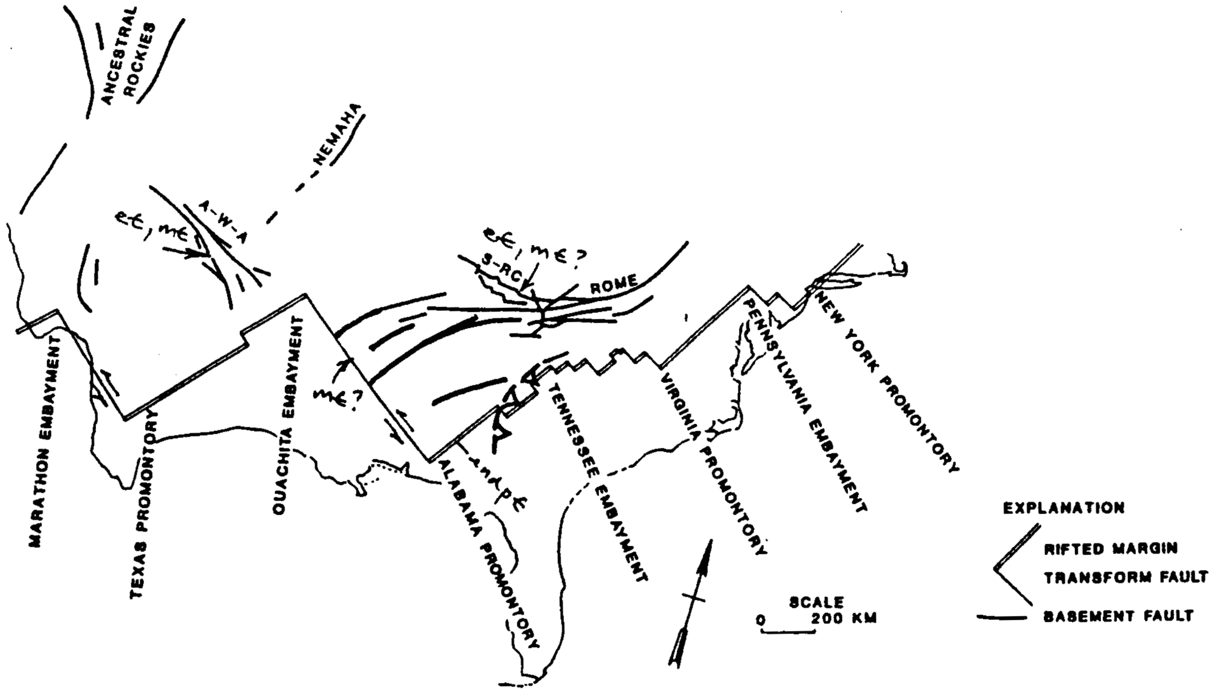

Figure 2: Early Phanerozoic edge of North America as a rift-transform fault succession from Thomas (1983); probable ages of last rifting and continent-ocean lithospheric boundary locally indicated; thrust symbols indicate southernmost recognized influence of Ordovician Taconian convergence.

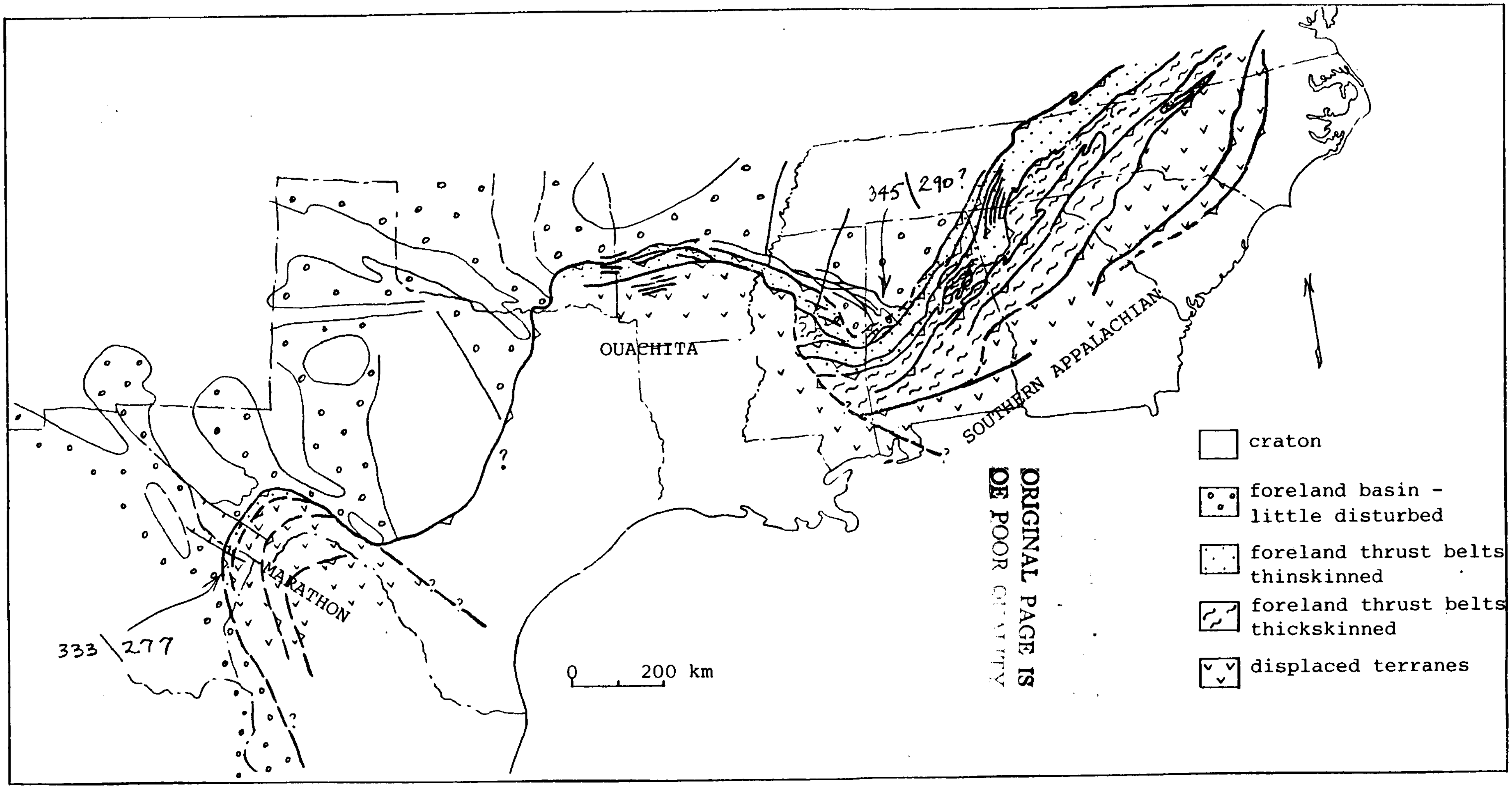

Figure 3: Late Paleozoic collision zone of southern North America; assumes Ouachita and Marathon allochthons are far-travelled with respect to North America; numbers indicate age of beginning/cessation of motion in displacement zone in ma.

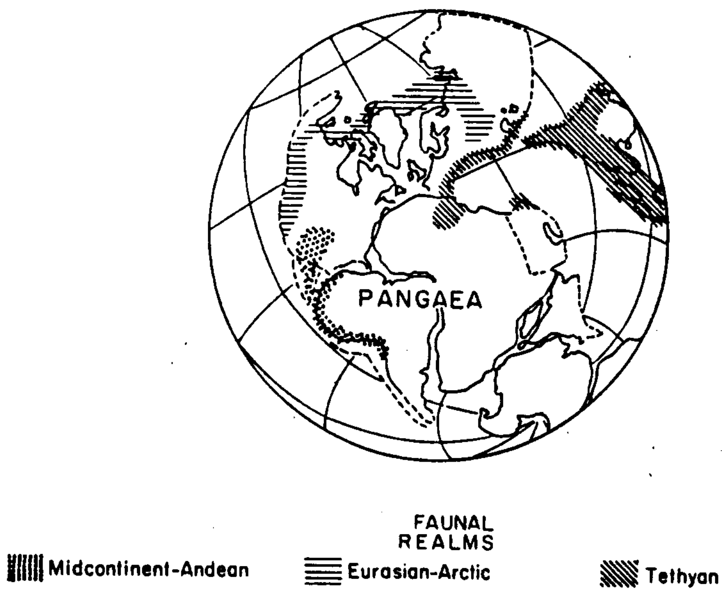

Figure 4: Pangea early in Triassic time; North America - South America fit according to faunal realms by Ross (1979).

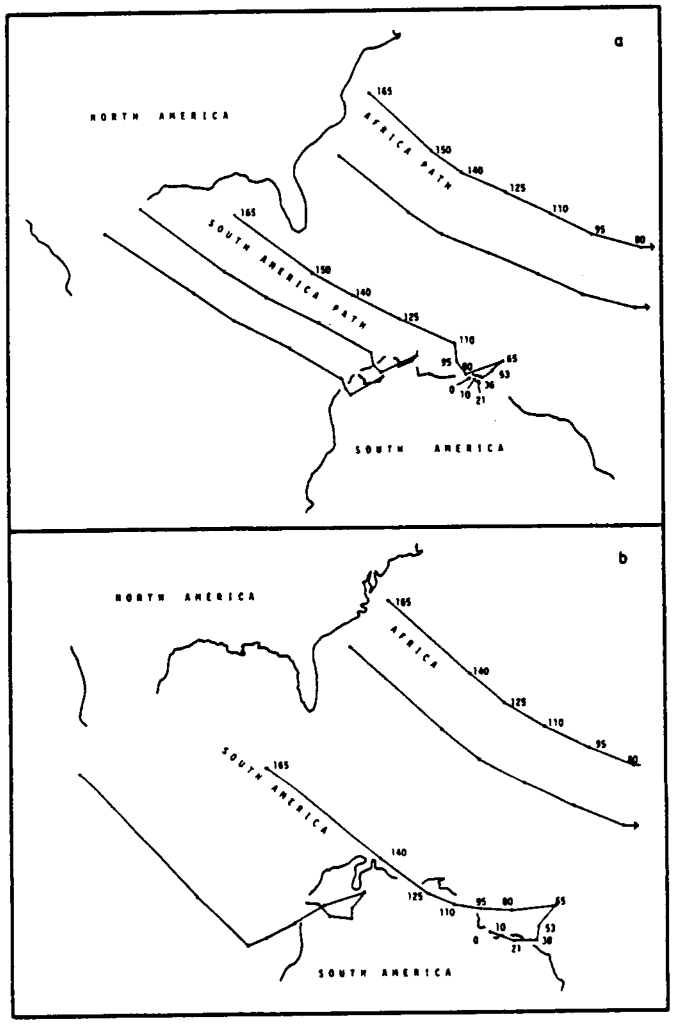

Figure 5: Mesozoic cratonal translation vector trajectories using NAm-Af and SAm-Af rotation poles;
a) Pindell and Dewey (1982);
b) Sclater and others (1977).

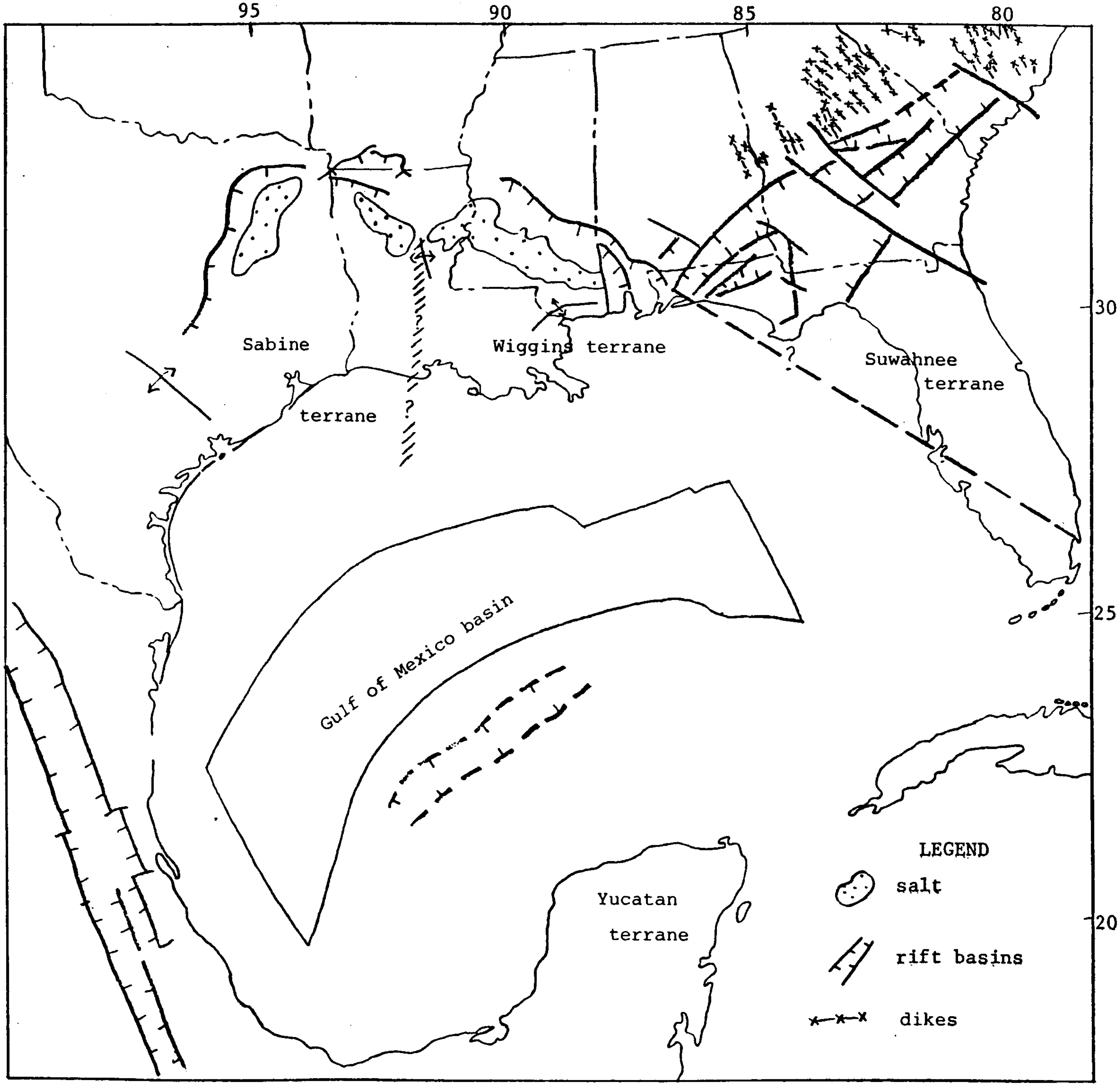

Figure 6: Rift-phase structures of the Gulf coastal region generated during Triassic-Jurassic extension within Pangea.

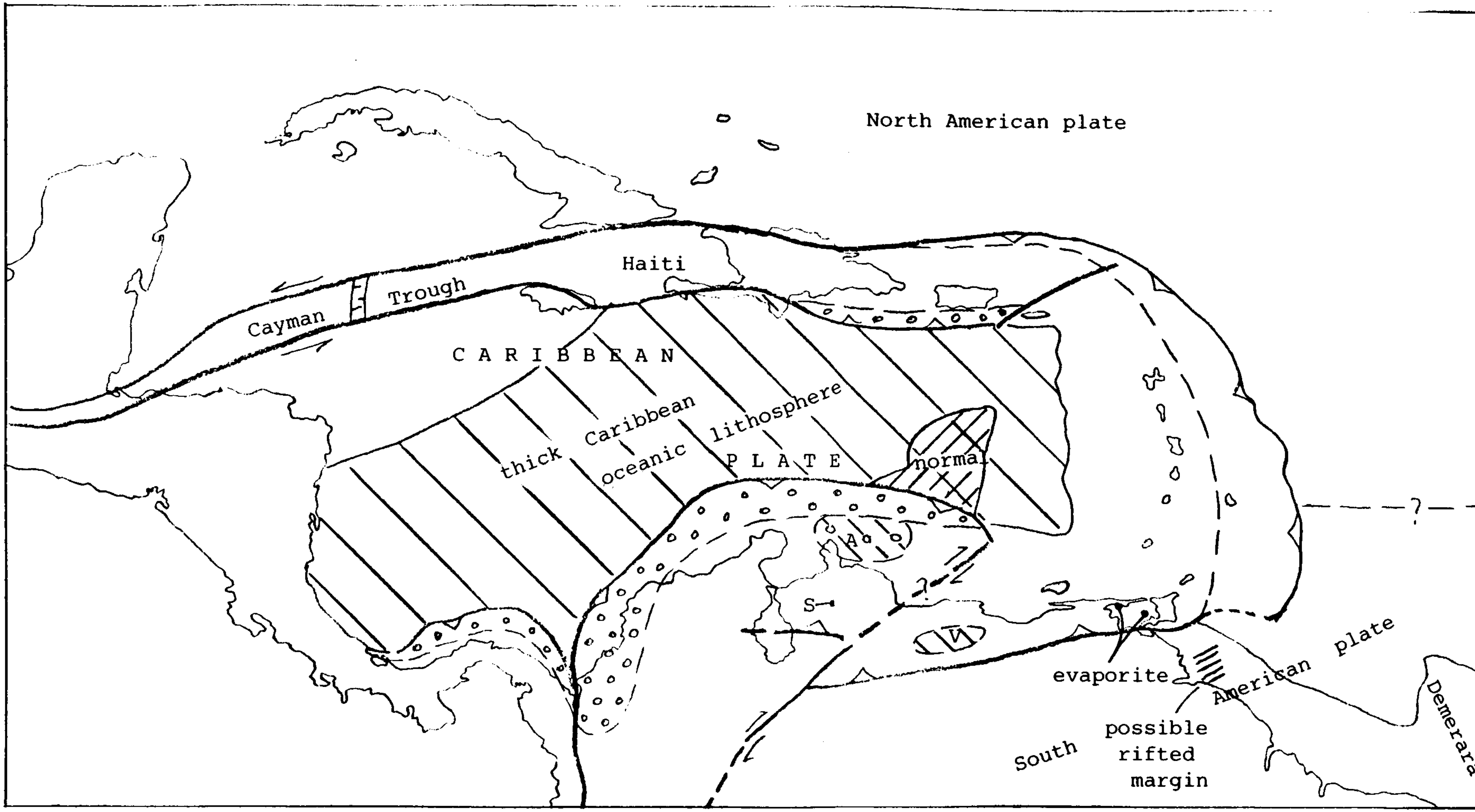

Figure 7: Caribbean region, showing sites of early Mesozoic features possibly related to rifting of Pangea and possible onland sites of Caribbean lithosphere. A=Netherlands Antilles; V=Villa de Cura; S=Siquisiqui; lined patterns=oceanic lithosphere; circles=sedimentary forearc, mainly Neogene.

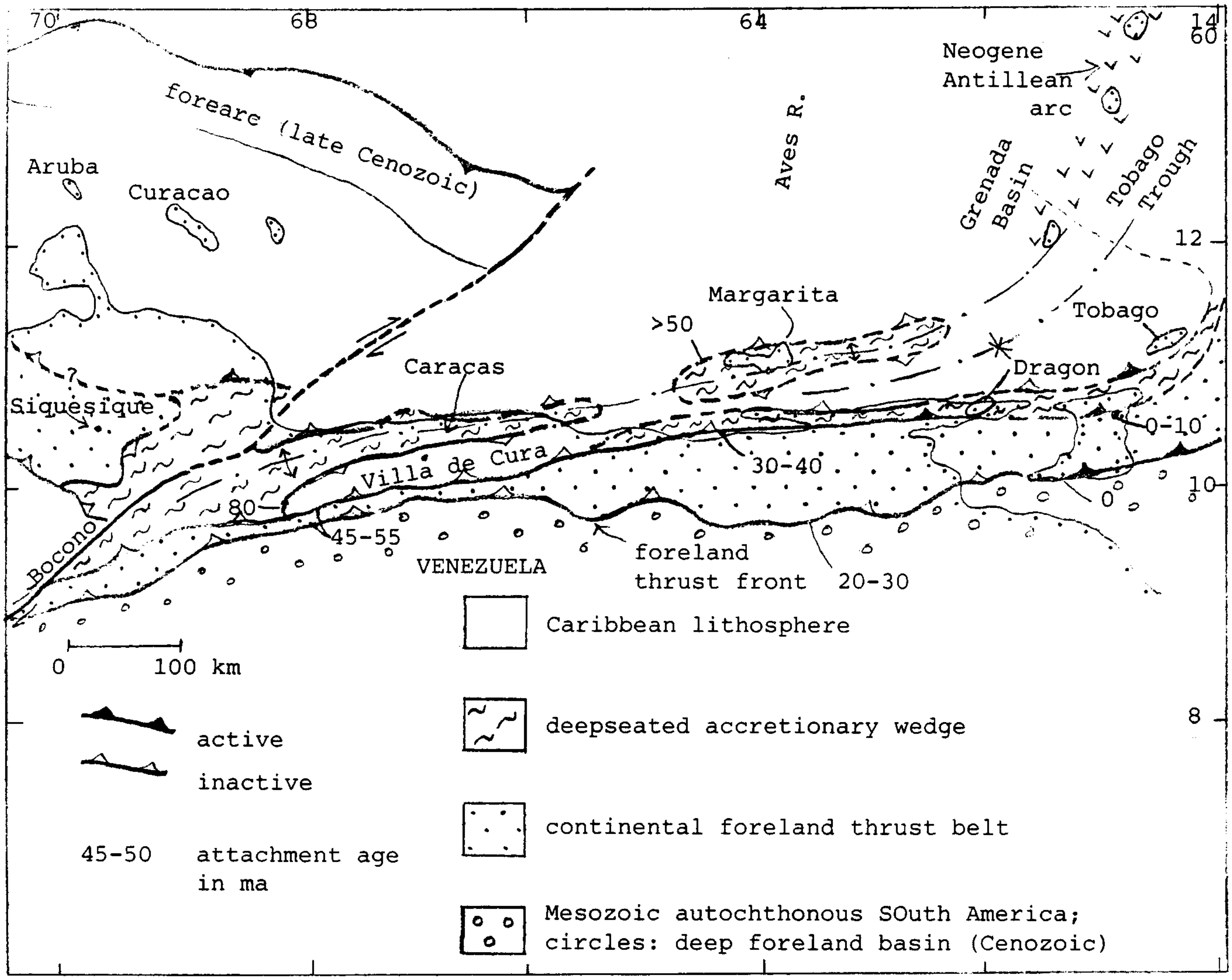

Figure 8: Late Cretaceous-Cenozoic collision belt of northern South America.

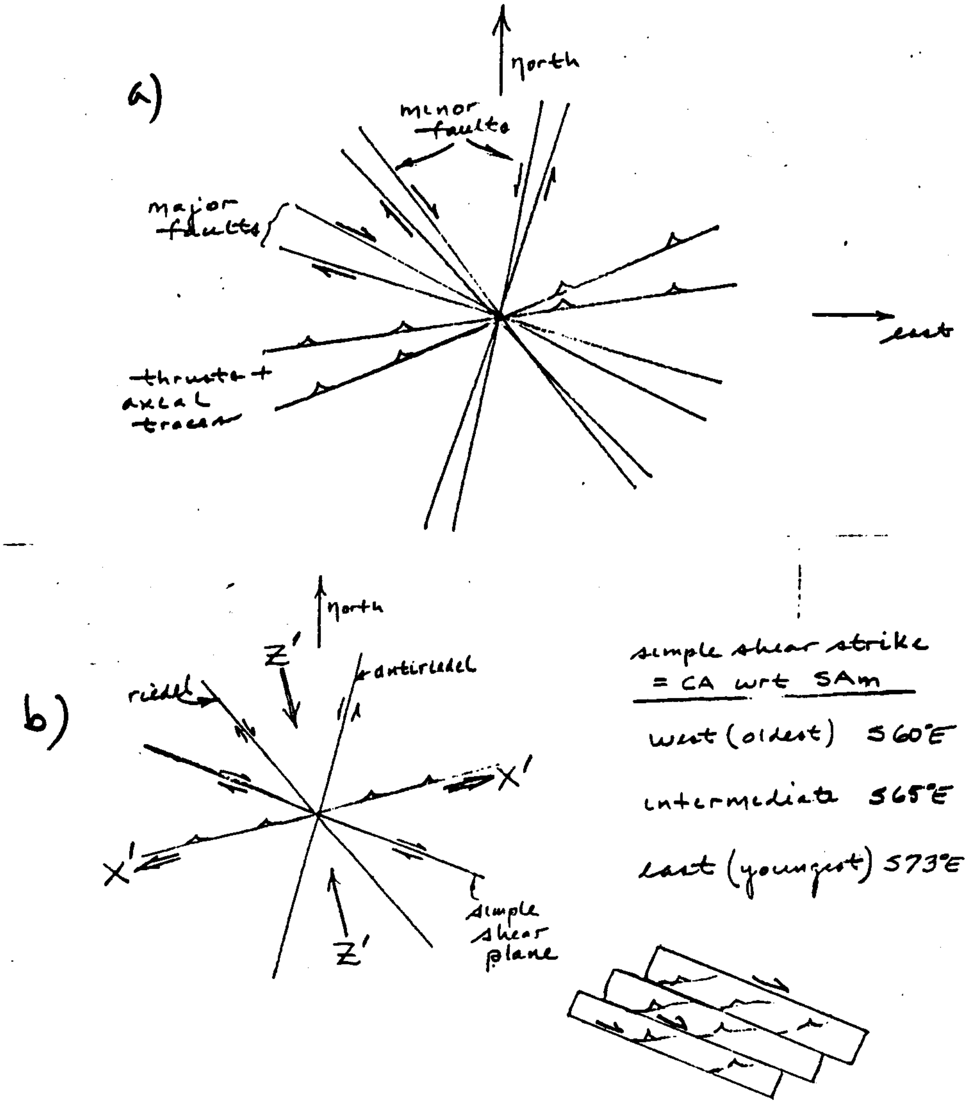

Figure 9: Orientation diagrams for eastern Venezuela foreland thrust belt; a) ranges of averaged strikes of structures in 5 domains along (EW) the belt from Cumana to Trinidad; b) interpretation of structures and horizontal stretch directions (x', z') and simple shear orientation.

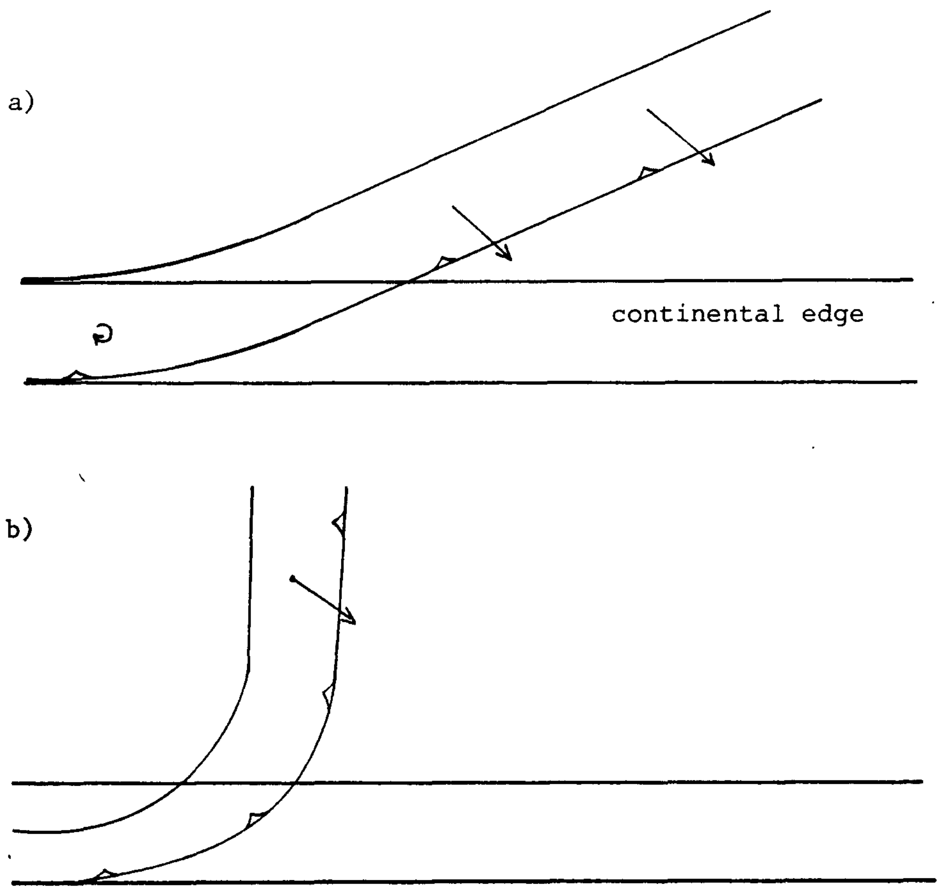

Figure 10: Alternative collision configurations for fringe of Caribbean lithosphere relative to the edge of northern South America; arrows show assumed plate convergence direction.

REFERENCES CITED

Angstadt, D. M., Austin, J. A., Jr., and Buffler, R. T., 1985, Early late Cretaceous to Holocene seismic stratigraphy and geologic history of Southeastern Gulf of Mexico: American Association of Petroleum Geologists Bulletin, v. 69, p. 977-995.

Beck, C., 1985, Las napas de Aragua-Cadena Caribe central y la historia mesozoica del margen Sur del Caribe a lo largo del meridiano de Caracas, *in* Caribbean Geodynamics, Mascle, A., et al., eds., Ed. Technip, p. 541-552.

Beck, M. E., Jr., 1986, Paleomagnetic data from active plate margins of South America. IGCP Symp. 1986:

Beets, D.J., Maresch, W.V., Klaver, G.T., Mottana, A., Bocchio, R., Beunk, F.F., and Monen, H.P., 1984, Magmatic rock series and high pressure metamorphism as constraints on the tectonics history of the southern Caribbean, *in* The Caribbean-South American Plate Boundary and Regional Tectonics, Geological Society of America Memoir 162, Bonini, W.E., ed., p. 95-130.

Bellizzia, A.G., 1972, Is the entire Caribbean mountain belt of Northern Venezuela allochthonous?, Studies in Earth and Space Science, Geological Society of America Memoir 132, p. 363-368.

—— 1985, Sistema montanosa del Caribe - una cordillera alóctona en la parte norte de America del Sur: Proc. 6th Congr. Geol. Venez., v. 10, p. 6657-6835.

Buffler, R., and 6 authors, 1983, Atlas 6: Gulf of Mexico; Ocean Margin Drilling Program: Woods Hole, Marine Science International.

—— 1988, Continent-Ocean Transect 14: Ouachita to Yucatan, DNAG: Geological Society of America.

Campos, Victor, 1981, Une transversale de la chaine caraibe et de la marge venezuelienne dans le secteur de Carupano (Venezuela oriental): Structure geologique et evolution geodynamique [Ph.D. thesis]: University de Bretagne Occidentale, 160 p.

Diebold, J.B., Stoffa, P.L., Buhl, P., and Truchan, M., 1981, Venezuela Basin crustal structure: Journal of Geophysical Research, v. 86, p. 7901-7923.

Donnelly, T.W., 1973, Magnetic anomaly observations in the eastern Caribbean Sea: Initial Reports of the Deep Sea Drilling Project, v. 15, p. 1023-1030.

Donnelly, T.W., Melson, W., Kay, R, and Rogers, J.J.W., 1973, Basalts and dolerites of Late Cretaceous age from the Central Caribbean: Initial Reports of the Deep Sea Drilling Project, v. 15, p. 989-1011.

Edgar, N.T., Saunders, J.B., Bolli, H.M., Boyce, R.B., Donnelly, T.W., Hay, W.W., Maurasse, F., Prell, W., Premoli-Silva, I., Riedel, W.R., and Schneidermann, N., 1973, Site 153: Initial Reports of the Deep Sea Drilling Project, v. 15, p. 367-406.

Ghosh, N., Hall, S.A., and Cassey, J.F., 1984, Sea floor spreading magnetic anomalies in the Venezuela Basin: Geological Society of America Memoir 162, p. 65-80.

Gonzales de Juana, C., Iturralde, J.M., and Picard, X., 1980, Geologia de Venezuela y de sus cuencas petroliferas, vol. I and II: Caracas, Ediciones Foninves, 1031 p.

Gose, W.Z., and Sanchez-Barreda, 1981, Paleomagnetic results from southern Mexico: Geofisica Internacional, v. 20, p. 163-175.

Hall, D.J., Cavanaugh, T.D., Watkins, J.S., and McMillen, K.J., 1982, The rotational origin of the Gulf of Mexico based on regional gravity data, *in* Studies in Continental Margin Geology, Watkins, J.S., and Drake, C.L., eds., American Assoc. of Petroleum Geologists Memoir 34, p. 115-126.

Kafka, A.L., and Weidner, D.J., 1979, The focal mechanisms and depths of small earthquakes as determined from Rayleigh-wave radiation patterns: Bulletin of the Seismological Society of America , v. 69, p. 1379-1390.

Keen, C., and Haworth, R., 1986, D-2 Transform margin south of Grand Banks: Offshore eastern Canada, Continent/Ocean Transect #2, Geol. Soc. of America, 6 p.

Klaver, G.T., 1987, The Curacao lava formation: an ophiolitic analog of the anomalous thick layer 2B of the mid-Cretaceous oceanic plateaus in the western Pacific and central Caribbean, GUA Papers of Geology, series 1, no. 27, 168 p.

Ladd, J. W., and Watkins, J. S., 1979, Tectonic development of trench-arc complexes on the northern and southern margins of the Venezuela Basin, *in* Geological and Geophysical Investigations of Continental Margins, Amer. Assoc. Petrol. Geol. Memoir 29, Watkins, J. S., Montadert, L., and Dickerson, P. W., eds., Tulsa, OK, AAPG, p. 363-371.

Larson, R.L., and Schlanger, S.O., 1981, Cretaceous volcanism and Jurassic magnetic anomalies in the Nauru basin, western Pacific Ocean: Geology, p. 480-484.

Lillie, R.J., Nelson, K.D., deVoogd, B., Brewer, J.A., Oliver, J.E., et al., 1983, Crustal structure of Ouachita Mountains, Arkansas: a model based on integration of COCORP reflection profiles and regional geophysical data: American Association of Petroleum Geologists Bulletin, v. 67, p. 907-931.

Macdonald, K. C., and Holcombe, T. L., 1978, Inversion of magnetic anomalies and sea-floor spreading in the Cayman Trough: Earth and Planetary Science Letters, v. 40, p. 407-414.

Maresch, W.V., 1972, The metamorphism and structure of NE Margarita, Venezuela [Ph.D. Thesis]: Princeton, NJ, Princeton University, 278 p.

—— 1974, Plate tectonic origin of the Caribbean Mountain System of northern South America: Discussion and proposal: Geological Society of America Bulletin, v. 85, p. 669-682.

Molnar, P., and Sykes, L.R., 1969, Tectonics of the Caribbean and Middle America regions from focal mechanisms and seismicity: Geological Society of America Bulletin, v. 80, p. 1639-1648.

Pardo, G., 1975, Geology of Cuba, *in* The Ocean Basins and Margins, v. 3, The Gulf of Mexico and the Caribbean, Nairn, A.E.M., and Stehli, F.G., eds., New York, Plenum Press, p. 553-615.

Perez, O.J., and Aggarwal, Y.P., 1981, Present-day tectonics of the southeastern Caribbean and northeastern Venezuela: Journal of Geophysical Research, v. 86, B11, p. 10791-10804.

Pindell, J.L., 1985, Alleghenian reconstruction and subsequent evolution of the Gulf of Mexico, Bahamas, and proto-Caribbean: Tectonics, v. 4, p. 1-39.

Pindell, J., and Dewey, J. F., 1982, Permo-Triassic reconstruction of western Pangea and the evolution of the Gulf of Mexico/Caribbean region: Tectonics, v. 1, p. 179-213.

Rabinowitz, P.D., and LaBrecque, J., 1979, The Mesozoic South Atlantic Ocean and evolution of the continental margins: Journal of Geophysical Research, v. 84, p. 5973-6002.

Rankin, D.W., 1975, The continental margin of eastern North America in the southern Appalachians: The opening and closing of the proto-Atlantic Ocean: American Journal of Science, v. 275-A, p. 298-336.

Robertson, P., Burke, K., and Wadge, G., 1986, Structure of the Melajo Clay near Arima, Trinidad, and strike-slip motion in the El Pilar Fault Zone: Trans. 1st Geol. Congress, 1985, p. 21-33.

Rosencrantz, E., Ross, M.I., and Sclater, J., 1988, The age and spreading history of the Cayman Trough as determined from depth, heatflow and magnetic anomalies: Journal of Geophysical Research, in press.

Ross, C.A., 1979, Late Paleozoic collision of North and South America: Geology, v. 7, p. 41-44.

—— 1986, Paleozoic evolution of southern margin of Permian basin: Geological Society of America Bulletin, v. 97, p. 536-554.

Salvador, A., 1987, Late Triassic-Jurassic paleogeography and origin of Gulf of Mexico Basin: American Association of Petroleum Geologists Bulletin, v. 71, p. 419-451.

Schlager, E., Buffler, R.T., Angstadt, D., and Phair, R., 1984, Geologic history of the southeastern Gulf of Mexico, *in* Init. Reports Deep Sea Drilling Project, v. 77, R.T. Buffler, W. Schlager, et al., eds., p. 715-738.

Schubert, C., 1981, Are the Venezuelan fault systems part of the southern Caribbean plate boundary?: Geol. Rundsc., v. 70, p. 542-551.

Sclater, J.B., Hellinger, S., and Tapscott, C., 1977, The paleobathymety of the Atlantic Ocean from the Jurassic to the present: Journal of Geology, v. 85, p. 509-552.

Sen, G., Hickey-Vargas, R., Waggoner, G., and Maurass, F., 1987, Geochemistry of basalts from the Dumisseau Formation, southern Haiti: Implications for the origin of the Caribbean Sea crust: Earth and Planetary Science Letters, in press.

Silver, E.A., Case, J.E., and MacGillavry, H.J., 1975, Geophysical study of the Venezuelan Borderland: Geological Society of America Bulletin, v. 86, p. 213-226.

Skerlec, G.M., and Hargraves, R.B., 1980, Tectonic significance of paleomagnetic data from northern Venezuela: Journal of Geophysical Research, v. 85, p. 5303-5315.

Speed, R. C., 1985, Cenozoic collision of the Lesser Antilles arc and continental South America and the origin of the El Pilar fault: Tectonics, v. 4, p. 41-69.

—— 1986, Cenozoic tectonics of the southeastern Caribbean and Trinidad, Proc. 1st Geol. Conf., Geol. Soc. T.&T. 1985, p. 270-280.

Stearns, C., Mauk, F.J., and van der Voo, R., 1982, Late Cretaceous-Early Tertiary paleomagnetism of Aruba and Bonaire (Netherland Leeward Antilles): Journal of Geophysical Research, v. 87, B2, p. 1127-1141.

Stephan, J.F., 1977, El contacto cadena Caribe-Andes Meridenos entre Carora y El Tocuyo (Edo. Lara): Observaciones en el Occidente Venezolano, Mem. Congr. Geol. Venez. 2: Min. Energia y Minas and Soc. Ven. de Geol., p. 789-817.

Stephan, J.F., Beck, C, Bellizzia, A., and Blanchet, R., 1980, La chaine caraibe du Pacifique a L'Atlantique, Bur. Recherches Geol. Min. Mem. 115 (Paris): p. 38-59.

Sykes, L.R., McCann, W.R., and Kafka, A.L., 1982, Motion of Caribbean plate during last 7 million years and implications for earlier Cenozoic movements: Journal of Geophysical Research, v. 87, p. 10656-10676.

Thomas, N.A., 1977, Evolution of Appalachian-Ouachita salients and recesses from reentrants and promontories in the continental margin: American Journal of Science, v. 277, p. 1233-1278.

Thomas, W.A., 1983, Continental margins, orgenic belts, and intracratonic structures: Geology, v. 11, p. 270-272.

—— 1985, The Appalachian-Ouachita connection: Paleozoic orogenic belt at the southern margin of North America: Annual Reviews of Earth and Planetary Science, v. 13, p. 175-199.

Ville, G.N., 1979, Geologic map and cross section, eastern Ouachita Mountains, Arkansas: GSA, v. MC-28F, p. 8.

Wadge, G., and Shepherd, J.B., 1984, Segmentation of the Lesser Antilles subduction zone: Earth and Planetary Science Letters, v. 71, p. 297-304.

Zietz, I. (Director of Compilation) 1982, Composite magnetic anomaly map of the United States: U.S. Geol. Survey.